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SOME THEORETICAL ASPECTS OF CONTAMINATION IN ACID MAGMAS

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ABSTRACT

Contamination in acid magmas is intimately connected with the volatiles which they contain. The chief function of these volatiles is to form a medium of low viscosity between the xenolith (or margin) and the magma through which material can diffuse with comparative freedom. This diffusion of substance to and from a xenolith constitutes reciprocal reaction. Its object is to stabilise the mineral assemblage of the xenolith so that its mineral phases are in equilibrium with the contaminated magma. The material gained by reciprocal reaction is incorporated in phases which crystallise from the contaminated magma. But material is also gained by the magma through other methods which may be loosely described as mechanical and which disintegrate the xenolith so that their constituent minerals are set free in the magma. So far, three such methods have been noted.

INTRODUCTION

Ever since petrology, as distinct from petrography, came into its own the rôle played by assimilation in the formation of igneous rocks has been the subject of repeated controversy. Some authors have maintained that vast quantities of material may be absorbed by a magma; others hold that it is a strictly local phenomenon. Some say that the resulting product will always be of abnormal composition or appearance; others reply that perfectly normal igneous rock types may result. In this way a large literature on the subject has developed, full of conflicting opinions and statements. Curiously enough, the point to which least attention has been paid is the means

by which the foreign material is incorporated. Almost the only contribution we have to this side of the subject is the work of N. L. Bowen.[1] This deals with the reactions which he considers take place when xenoliths are immersed in magma and helps to explain how and why the xenoliths and their enclosing magma are modified. The other writers who have directed serious attention to this aspect of the problem are A. Harker[2] and H. H. Read,[3] the latter, however, dealing with a basic magma.

In the following pages an attempt is made to show what happens, according to the writer's view, when foreign material is added to an acid magma and the various ways in which this material is incorporated. It should be emphasised that in the present contribution acid magmas only are considered and, moreover, only plutonic acid magmas.

THE QUESTION OF SUPERHEAT

One of the greatest difficulties which beset the early advocates of assimilation was the question of superheat in the magma. Their opponents were quick to point out that in order to dissolve any appreciable amount of foreign material the magma must possess a large amount of superheat, a condition not likely to be fulfilled, at any rate in acid magmas. Various answers were made to this, but none of them can be accounted very satisfactory. In dealing with this question it was assumed that all the foreign material passed into solution in the magma and then the modified magma solidified, thus producing the hybrid rock.

It was left to Bowen[4] to show that superheat is required only to bring the foreign material up to the temperature of the magma. After that, what he terms "reactive precipitation" plays the dominant part. As the question is fully dealt with by him in the fore-

[1] N. L. Bowen, "The Behavior of Inclusions in Igneous Magmas," *Jour. Geol.*, Vol. XXX (1922), Suppl., pp. 513–70; *Evolution of the Igneous Rocks* (Princeton, 1928), chap. x.

[2] A. Harker, "Igneous Rock Series and Mixed Igneous Rocks," *Jour. Geol.*, Vol. VIII (1900), pp. 389–99; *Natural History of the Igneous Rocks* (London, 1909), pp. 335–36; "Tertiary Igneous Rocks of Skye," *Mem. Geol. Surv. U.K.* (1904), pp. 231–32.

[3] H. H. Read, "The Petrology of the Arnage District," *Quar. Jour. Geol. Soc.*, Vol. LXXIX (1923), pp. 479–84.

[4] N. L. Bowen, *Evolution of the Igneous Rocks* (Princeton, 1928), pp. 182–97.

going reference, it is not necessary to say much here. The main point to be noticed for our purpose is that if the xenolith is composed of material earlier in the reaction series than that with which the magma is saturated, then the magma cannot dissolve the xenolith, but will react with it to produce, as far as possible, that member of a reaction series with which it is itself saturated. This type of reaction is exothermic and consequently no superheat is required for its accomplishment. The only time material is actually dissolved by the magma is when the xenolith is composed of a member of a reaction series later than that member of the series with which the magma is saturated. This type of reaction is endothermic and thus would not proceed far without a supply of heat. This need not be magmatic superheat, however. It can be heat of crystallisation as Bowen has pointed out. As we are dealing here with acid magmas, the components of which tend to be late in the various reaction series, the endothermic type of reaction will, in general, be unimportant and most of the reactions which take place during the contamination will be exothermic.

THE IMPORTANCE OF THE VOLATILE CONSTITUENTS

The important part which the volatile constituents play during contamination in acid magmas has been emphasised before and the idea expressed that there was a preferential absorption of the volatile constituents by the xenoliths.[5] This conclusion was based partly on the relative abundance of apatite found in the xenoliths and more contaminated portions of the magma when compared with the amount present in the invaded or invading rock. This apatite is characteristically present as thin needles penetrating the other minerals and often of exceedingly small dimensions. It becomes a matter of interest, therefore, to see if this relative abundance of apatite has been noticed in other examples. On examining the literature one does, indeed, find many such cases.

Harker[6] appears to have been the first to notice and put on record the relative abundance of apatite in hybrid rocks when dealing with

[5] S. R. Nockolds, "The Dhoon (Isle of Man) Granite; a Study in Contamination," *Min. Mag.*, Vol. XXII (1931), p. 506.

[6] "Carrock Fell: A Study in the Variation of Igneous Rock Masses," Part II, *Quar. Jour. Geol. Soc.*, Vol. LI (1895), pp. 125–39.

the Carrock Fell granophyre. In 1904 he[7] again recorded a number of instances of hybrid rocks from Skye carrying an abundance of apatite. It is particularly significant to note that in marscoite, a highly contaminated rock, a striking feature is "the immense number of minute apatite needles which it contains." Rastall[8] mentions apatite as an abundant accessory of the "mixture rocks" produced at Burtness Combe and Scrivenor[9] describes basic patches in a granite as containing "abundant apatite." In dealing with hybrid rocks produced by the action of pegmatite on gabbro, De Lapparent[10] finds that apatite is lacking in the pegmatite and in the gabbro, but is abundant in the altered xenoliths derived from the gabbro.

Turning to more recent years, we find that Grantham[11] in his paper on the Shap granite, mentions the abundance of apatite needles in the andesitic inclusions and in the early basic granite (a hybrid). A similar concentration in a hybrid has been observed in a composite dyke at Klipsriversberg.[12] Wells and Wooldridge[13] find an abundance of apatite in basic xenoliths enclosed by granite. In the recently published memoir on Ardnamurchan more examples may be found.[14] A particularly interesting one is the tonalite (a hybrid) of the Interior Complex of Centre 3 Ring Dykes.[15] Where this rock is extremely acid (i.e., little contaminated) it is stated that ". . . . there is no strong concentration of apatite such as is a feature of the more definitely tonalitic mass."[16] Abundant needles of apatite have

[7] "Tertiary Igneous Rocks of Skye," *op. cit.*, pp. 172, 181, 183, 186, 195.

[8] R. H. Rastall, "The Buttermere and Ennerdale Granophyre," *Quar. Jour. Geol. Soc.*, Vol. LXII (1906), p. 263.

[9] J. B. Scrivenor, "Igneous Rocks of Singapore," *Geol. Mag.*, N.S., Vol. VI (1909), p. 21.

[10] J. de Lapparent, "Les Enclaves des granite du Mont Haya en pays basque," *Bull. Soc. Min. France*, Vol. XXXVI (1913), p. 169.

[11] D. R. Grantham, "The Petrology of the Shap Granite," *Proc. Geol. Assoc. London*, Vol. XXXIX (1928), p. 311.

[12] D. P. McDonald, "Note on a Composite Dyke in the Klipsriversberg," *Trans. Geol. Soc. S. Africa*, Vol. XV (1912), p. 140.

[13] A. K. Wells and S. W. Wooldridge, "The Rock Groups of Jersey," *Proc. Geol. Assoc. London*, Vol. XLII (1931), p. 204.

[14] "Geology of Ardnamurchan, N.W. Mull, and Coll," *Mem. Geol. Surv. Scotland* (1930), p. 279.

[15] *Ibid.*, p. 336.

[16] *Ibid.*, p. 340.

been noted in connection with the xenoliths in the Bibette Head (Alderney) granite.[17] In the Ach'uaine hybrids apatite is abundant though occurring only in small amount in the granitic end member.[18] Several other examples could be given.

Besides this question of the apatite, however, there are other lines of evidence pointing to the important part played by the volatile constituents. For instance, although the granophyres of Skye, Mull, Ardnamurchan, etc. (in West Scotland), are usually augite bearing, yet, when they are contaminated, they develop hornblende and even biotite. Indeed, the authors of the Mull memoir say,[19] "hornblende —or biotite—granophyres are very rare in Mull, except as a product of contamination."

All the examples chosen thus far have been cases where the xenoliths are of basic igneous origin. There are, however, indications that the same relative concentration of volatiles takes place where sedimentary rocks are invaded. The evidence is more difficult to gather in these rocks because apatite, in general, ceases to be a guide, owing to the different nature of the materials. An interesting example has been recorded by Grantham[20] where a calc-silicate rock immersed in granite has developed zones which show quite clearly the part played by volatiles, especially water and sulphur, during their formation. The most highly altered zone comes, in the end, to resemble the granite itself. In the granophyre-sandstone assimilation zone of the Glas Bheinn granophyre,[21] the granophyre, which usually carries augite and some hornblende, carries hornblende and biotite where much material has been incorporated. In the Loch Ailsh intrusion,[22] dolomite xenoliths enclosed in perthosite have developed abundant biotite and the contaminated rock formed by interaction contains apatite and pyrite, the normal perthosite being free from these min-

[17] Nockolds, "The Contaminated Granite of Bibette Head, Alderney," *Geol. Mag.*, Vol. LXIX (1932), p. 442.

[18] "The Geology of Strath Oykell and Lower Loch Shin," *Mem. Geol. Surv. Scot., Expl. Sheet 102* (1926), pp. 157–66.

[19] "Tertiary and Post-tertiary Geology of Mull," *Mem. Geol. Surv. Scot.* (1924), p. 21.

[20] D. R. Grantham, *op. cit.*, pp. 318–19.

[21] "Tertiary and Post-tertiary Geology of Mull, etc.," *op. cit.*, pp. 169–70.

[22] "Geology of Strath Oykell and Lower Loch Shin," *op. cit.*, pp. 95–98.

erals. Finally, the contaminated rock resulting from the incorporation of iron formation by granite in the Onaping map area (Canada) contains "large rods of apatite" which are not present in the normal granite.[23]

The foregoing instances are, perhaps, sufficient to show that this apparent concentration of volatiles in xenoliths and in the more contaminated portions of acid magmas is a factor of real significance and that the hypothesis does not rest on one or two isolated occurrences, but on a considerable body of evidence.

It will be noticed in the foregoing pages that the phrase "relative concentration" has been used throughout. It is believed that there has never been an actual concentration of volatiles in the xenoliths over and above that contained in the surrounding acid magma. The relative concentration is produced because portions of the volatiles are fixed in solid phases in the xenolith. In the final product there is a concentration of volatile material, but this does not mean an actual concentration of the volatiles during any one period of the alteration of the xenolith. Similar considerations apply to the more contaminated portions of the magma. Part of the apatite here may be gained by mechanical disintegration of the xenoliths, part by the entry of lime into the magma from the xenoliths causing a precipitation of apatite.

One of the most striking features of xenoliths is the way in which they retain their shape even though they may be completely altered. It has been commented on by the authors of the Mull memoir,[24] by Adams and Barlow,[25] and by Harker.[26] This preservation of outline shows that the xenoliths remain essentially solid throughout the whole course of their alteration. There has been no true solution, but the alteration has been one of molecular replacement, a minute portion of one of the constituents going into solution at any one moment and being replaced by new material. This necessitates free diffusion into the xenolith and also out of it. The foregoing does not,

[23] W. H. Collins, "Onaping Map-Area," *Geol. Surv. Can. Mem. 95* (1917), p. 56.

[24] "Tertiary and Post-tertiary Geology of Mull," *op. cit.*, p. 354.

[25] F. D. Adams and A. E. Barlow, "Geology of the Haliburton and Bancroft Areas, Province of Ontario," *Geol. Surv. Can. Mem. 6* (1910), p. 109.

[26] "Geology of the Small Isles," *Mem. Geol. Surv. Scot.* (1908), p. 112.

of course, apply to xenoliths more acid than their enclosing magma, but as we are dealing with acid magmas, the xenoliths will usually be more basic.

Thus, there is good evidence from the rocks themselves of (*a*) free diffusion of material into the xenoliths causing their alteration, (*b*) the entry of volatile constituents into the xenoliths from the magma, and (*c*) the alteration of the xenoliths after the volatiles have entered them. Is there any causal connection between these facts? Are the volatiles responsible, in some way, for this free diffusion of magma material into the xenoliths?

Of the volatiles present in an acid magma, water is undoubtedly the most important. The work of Niggli,[27] Morey and Niggli,[28] and Morey[29] has shown that we must regard the water of a magma as being in the liquid or vapour state and not in the state of a gas. In the case of plutonic magmas, the water will be all or nearly all in the liquid state, owing to the high external pressure. Of course, if the external pressure is less than the internal pressure of the magmas, then some of the water will appear in the vapour state, the amount depending on the pressure difference. As, however, the maximum vapour pressure of a silicate-water system is only attained after a considerable amount of crystallisation has taken place, the most likely period for the formation of this vapour phase will be at a time when contamination and reaction cannot take place, owing to advancing crystallisation.

To all intents and purposes, therefore, the diffusion of the volatiles (mainly water) into a xenolith means diffusion of a liquid. To reach the xenolith, this liquid has to diffuse through the magma and then enter the xenolith where true diffusion is largely overshadowed by capillary and subcapillary flow. Diffusion in liquid magmas composed predominantly of silicate matter receives but scant support from Bowen who argues that the high viscosity would prevent it occurring through significant distances. It has already been shown,

[27] P. Niggli, "Die Gase in Magma," *Centrl. für Min. etc.* (1912), pp. 321–28.

[28] G. W. Morey and P. Niggli, "The Hydrothermal Formation of Silicates," *Jour. Amer. Chem. Soc.*, Vol. XXV (1913), pp. 1086–1130.

[29] G. W. Morey, "The Temperatures of Hot Springs and the Sources of Their Heat and Water Supply, II," *Jour. Geol.*, Vol. XXXII (1924), pp. 294–95.

however, that free diffusion of material into xenoliths does occur and so diffusion in the liquid state must take place through significant distances in order to keep up the supply of material necessary for the alteration of the xenoliths. It seems probable that Bowen lays too much stress on the viscosity. Little is known about the viscosity of mixtures whose individual viscosities are very different, but what is known points to the fact that a small proportion of the more fluid substance exercises a very large effect in reducing the viscosity of the mixture as a whole.[30] Temperature, of course, also reduces viscosity, but this is probably more than counterbalanced by pressure. The rate of change of viscosity with pressure for most liquids is large and, moreover, the more complex the molecules of the liquid, the greater the change. Water is an exception to the rule, the rate of change of viscosity being small.[31] So again, under the conditions existing in the magma, the influence of water would be very great in spite of its comparatively small concentration. It seems highly probable, therefore, that natural magmas are not so viscous as is sometimes made out and that diffusion through significant distances can take place, given time, with comparative ease.

When it comes to the question of the rate of diffusion through the liquid, we find that the volatiles and such simple silicates and other salts as tend to move with them have a very definite advantage over the complex silicate molecules of the magma. In the first place, the resistance to motion will be far less for the molecules of the volatiles and of the simple salts owing to their small molecular volume. Further, the mean free path of these molecules will be correspondingly greater than that of the large molecules of the complex silicates. Finally, although when several substances are diffusing each one diffuses at its own specific rate, yet it appears that the difference between their rates of diffusion, when alone, is somewhat increased when they diffuse simultaneously. This would tend to enhance the difference in diffusivity already existing between the complex silicate molecules and those of the volatiles and the simple salts. The cause of the diffusion is, of course, a concentration gradient. In the marginal rocks and in the xenoliths, the available concentration of the

[30] E. Hatschek, *The Viscosity of Liquids* (London, 1928), p. 162.

[31] P. W. Bridgman, *The Physics of High Pressure* (London, 1931), pp. 330–56.

diffusing molecules is, in the first place, almost nil. Once the volatiles have entered, however, diffusion into the xenolith will take place only when a portion of the volatiles is fixed in solid phases and the concentration thus reduced. In the case of the simple salts which, at the beginning, travel into the xenolith with the volatiles, diffusion will continue as long as they are being used up to effect the alteration of the xenolith. This subject is taken up in greater detail later on when it is discussed further under the heading of "Reciprocal Reaction."

Summing up the results just obtained, we may say that from a theoretical point of view there are indications that diffusion into xenoliths should be accomplished by the volatile components of a magma and the simple salts which tend to travel with them rather than by the complex silicate components. Whichever way we look at the matter we seem to reach the same conclusion. Evidence from the xenoliths themselves points to the fact that volatiles from the magma have entered them and further that their alteration has been one of molecular replacement with free diffusion of material from the enclosing magma. This molecular replacement must have been carried out either by the simple salts associated with the volatiles or by the complex silicate components of the magma. Theoretically, it would appear that, of the two, the simple salts are the only substances which would be likely to diffuse with sufficient freedom. Therefore it must be these which effect the alteration of the xenolith.

If the volatiles, carrying dissolved substances with them which effect the alteration of the xenolith, diffuse into the xenolith, then the elements displaced by this alteration must diffuse out into the magma. This naturally leads to the subject of reciprocal reaction.

RECIPROCAL REACTION

Reciprocal reaction was first defined by Read[32] working with sedimentary xenoliths in a basic magma, though a rather similar idea had been previously put forward by Johnston Lavis[33] and the same

[32] *Op. cit.;* "On Certain Xenoliths Associated with the Contaminated Rocks of the Huntly Mass, Aberdeenshire," *Geol. Mag.*, Vol. LXI (1924), pp. 441–42.

[33] H. J. Johnston Lavis, *Natural Sci.*, Vol. IV (1894), pp. 134–40.

thing had been hinted at by Harker[34] and De Lapparent.[35] It is the name given to the process whereby xenoliths and contact rocks gain certain constituents from the magma, giving other constituents to the magma in exchange.

It has already been pointed out that in this transference of material, lime, magnesia, and ferrous iron tend to be associated on the one hand, and potash, soda, and sometimes alumina, on the other.[36] Alumina, however, appears to be associated with either set of oxides on occasion. On looking through the literature of xenoliths enclosed in acid magmas it is possible to gain some further insight into the changes which take place. It may be noticed, for instance, that when soda is being transferred it is fixed predominantly by the lime in the xenolith.[37] It has been shown experimentally by Endell[38] that soda transfuses more rapidly than potash into a lime-rich nucleus—a fact which can also be noticed in some limestone xenoliths.[39] Similarly, if the xenolith is comparatively rich in magnesia, then potash transfuses in greater quantity than soda. Examples of this are found in the dolomitic xenoliths of the Loch Ailsh intrusion.[40] An interesting case in which potash has entered a xenolith in large quantity and where soda and alumina have actually been lost (in addition to lime) has been described by Jaskolski, an instance in which an amphibolite inclusion was immersed in granite.[41] Although potash, for instance, rather than soda tends to enter a xenolith rich in magnesia, yet this does not mean that potash can enter only xenoliths comparatively rich in magnesia. Similarly, soda does not enter only xenoliths rich in lime. Each oxide acts quite independently under different circum-

[34] "Tertiary Igneous Rocks of Skye," *op. cit.*, p. 183.

[35] *Op. cit.*, p. 159.

[36] Nockolds, "The Dhoon (Isle of Man) Granite," *op. cit.*

[37] See also A. Holmes, "Pre-Cambrian and Associated Rocks of Mozambique," *Quar. Jour. Geol. Soc.*, Vol. LXXIV (1918), p. 60.

[38] K. Endell, "Über Diffusionserscheinungen in Silikatschmelzen bei höheren Temeraturen," *Neue Jahr. für Min. etc.*, No. 2 (1913), pp. 129–54.

[39] W. G. Foye, "Nephelite Syenites of Haliburton County, Ontario," *Amer. Jour. Sci.*, Ser. 4, Vol. XL (1915), p. 433.

[40] "Geology of Strath Oykell and Lower Loch Shin," *op. cit.*, pp. 96–97.

[41] S. Jaskolski, "Les Amphibolites des Monts Tatra et leur Origine," *Acad. Polonaise Bull. Sci.*, Mat. et Nat. Ser. A. (1924), pp. 101–5.

stances, but one can say, with a fair amount of confidence, that if the xenolith is rich in magnesia then potash tends to enter rather than soda, whereas if the xenolith is rich in lime, soda tends to enter rather than potash. An example, which is apparently contradictory to what has been said above, has been cited by Eckermann.[42] His analyses show that potash rather than soda has entered limestone in contact with the magma. But the mineral assemblage found in the limestone at the contact points to the fact that this limestone was rich in magnesia, in which case the potash would be expected to transfuse. Eckermann maintains that the limestone was originally almost pure and the magnesia was also introduced by the magma. With this we cannot agree, especially as the magma was of pegmatitic nature. The entry of titanium into lime-bearing xenoliths is also a common phenomenon. Apparently the entry of titanium is also independent of the entry of any other oxide.

Moreover, in xenoliths immersed in acid magmas, the content of any of the oxides soda, potash, alumina, or titanium may exceed that found in either the parent-rock or the invading magma (as seen in the solid state). That is to say, the xenolith does not acquire, or attempt to acquire, a chemical composition intermediate between that of its parent and the invading magma. These reactions, which at first sight seem to have no meaning and to be purely capricious in their workings, acquire a new significance when considered from a point of view different from that of the chemical one. The magma has been primarily concerned in attaining phase equilibrium within the xenolith, producing solid phases which are in equilibrium with the magma as far as possible. Reciprocal reaction does not take place because the chemical composition of the xenoliths differs from that of the magma, but because their phasal composition is not in equilibrium under the prevailing conditions. A simple example will make this clear. Suppose we have a normal biotite granite magma and introduce a schist xenolith composed of biotite, orthoclase, and quartz; there will be no reciprocal reaction because the phases are stable under the conditions prevailing in the magma. Nevertheless, as regards chemical composition, the proportions of the oxides in

[42] H. von Eckermann, "The Rocks and Contact Minerals of the Mansjö Mountain," *Geol. Foren. Forhandl.*, Vol. XLIV (1922), p. 326.

the xenolith and in the granite may be very different. This does not mean that a biotite granite magma cannot be contaminated by a biotite schist, because, as will be seen later, reciprocal reaction is not the only way in which a magma may be contaminated.

It follows from this, of course, that the nearer the mineral composition of the xenolith to that of the enclosing magma, the easier it is for the assemblage to be stabilised. Thus, a basic igneous xenolith would stand more chance of becoming completely stabilised under a given set of conditions than, say, a limestone or calc-silicate xenolith. Similarly, a biotite schist would be more favourably situated than a highly aluminous sediment. Consequently we find that such minerals as corundum, andalusite, garnet, etc., in aluminous xenoliths and wollastonite, grossular, etc., in calcareous xenoliths still commonly persist. This is due only to the fact that stable equilibrium could not be reached under the prevailing conditions and such assemblages in xenoliths enclosed by acid magmas must be regarded as unstable. We agree entirely with Bowen when he says that such minerals "should be temporary or should survive only because of exhaustion of the liquid."[43] Nevertheless, these unstable assemblages are of great importance as they may sometimes show the way in which reaction proceeds.

The disappearance of unstable minerals in xenoliths during reciprocal reaction has been noted by Read.[44] He found the production of thick coatings of pale mica around andalusite, sillimanite, garnet, and staurolite, relics of these minerals often being left in the centre. The fact that the transformation of a xenolith often takes place in stages is well brought out by the calc-silicate inclusion described by Grantham.[45] The inner zone shows quartz, diopside, sericite, and doubtful wollastonite. The succeeding zones show, in order: (1) quartz, sericite, stellate clusters of wollastonite, diopside, and incipient chlorite; (2) quartz, sericite, and chlorite; (3) quartz, pale biotite, a little sericite, and well-developed pyrite; and (4) a marginal zone composed of almost equal quantities of orthoclase and quartz together

[43] *Op. cit.*, p. 216.

[44] "The Igneous and Metamorphic History of Cromar, Deeside, Aberdeenshire," *Trans. Roy. Soc. Edinburgh*, Vol. LV (1927), pp. 323–33.

[45] *Op. cit.*, p. 319.

with biotite and a little plagioclase. This last zone thus comes to resemble the enclosing granite.

Once phase equilibrium has been attained, reciprocal reaction will cease. It has already been noted that the volatiles enter the xenoliths from the magma, and it must be they, or rather the substances which they carry with them, that effect the alteration of the xenoliths and allow reciprocal reaction to take place. In acid magmas this means essentially an introduction of silica, soda, potash, alumina, titanium, and sometimes iron, any or all of these elements being introduced in various examples. There are many ways in which the elements to be introduced could combine with the volatiles or with silica or with the hydroxide radical so as to be soluble in water, and the fact that they are carried in solution in water will favour their ionisation and consequently enhance their chemical activity and their diffusive power.

It does not matter whether the amount carried is great or small as long as some is carried. The reason for this is that the substances carried are fixed in solid phases in the xenolith and consequently removed from the system. This lowers the effective concentration, and so another small amount of the elements from the complex silicate portion of the magma will go into solution in the volatiles only to be fixed again, in its turn, when it enters the xenolith. Thus the process will go on until the solid phases of the xenolith are in equilibrium or until diffusion of material stops, owing to advancing crystallisation in the magma. The bases diffusing out from the xenolith as the result of reaction must also be in solution in the volatiles (water). The elements taken from the complex silicate portion of the magma are replaced by the elements which come from the xenolith. A nicely balanced state of equilibrium is maintained, and the volatiles themselves act as a means of communication between the silicate portion of the magma and the xenolith.

The diffusion of elements such as potash, soda, etc., into the xenolith is thus independent, in a way, of the diffusion of the volatiles. The volatiles (mainly water) diffuse through the magma, but once they are inside the xenolith their progress is guided by capillary and subcapillary flow. The elements which enter the xenolith, however, are subject to true diffusion all the time as they diffuse through

the magma and also through the main volatile constituent (water) which has entered the xenolith. All the bases of the complex silicate molecules of the magma will be represented, according to their solubilities, in the more volatile portion which enters the xenolith. The magma cannot be regarded as a simple mutual solution of complex silicates and volatiles (mainly water), because, at the temperatures existing in the magma, the volatiles are capable of attacking the complex silicates and dissociating them to a greater or lesser extent. Owing to this effect of the volatiles, balanced reactions will be present between the complex molecules and their dissociation products. The amount of the dissociation products formed will depend on the concentration of the volatiles and the solubility of the products themselves. We must imagine all the complex silicate molecules affected to some extent by this process.

When the volatiles diffuse into a xenolith they will carry a certain amount of these dissociation products with them. Reciprocal reaction then commences to take place. Certain of these substances are abstracted by the xenolith in its effort to gain a mineralogical constitution similar to that which is potentially present in the magma. At the same time the xenolith gives other elements to the volatiles representing the material replaced by the constituents which it has abstracted. The result is that the whole equilibrium existing between the complex silicate molecules of the magma and their dissociation products is upset. Suppose, for instance, that soda is being abstracted by the xenolith from its solution in the volatiles. At the beginning there was equilibrium between the amount of soda carried in solution in the volatiles and the amount present in the complex silicate molecules. When this soda is abstracted by the xenolith and precipitated in solid phases, it is removed from the system and thus the concentration in solution is reduced. This upsets the equilibrium existing between its concentration in the volatiles and in the complex silicate portion of the magma. Therefore some passes from the complex silicate portion to the volatiles, i.e., further dissociation takes place. At the same time, however, this soda is replacing some element, e.g., lime, in the xenolith. This replaced lime must go into solution in the volatiles in order to be removed from the xenolith. It cannot do this without upsetting the equilibrium already existing

between the lime compounds in solution in the volatiles and the complex lime silicates of the magma. The effect of this lime, therefore, is to decrease the amount of dissociation and consequently the complex silicate portion of the magma has its lime content increased. We may represent the sequence of events for the example which we have just taken diagrammatically as shown:

1. At first we have equilibrium like this:

Complex silicates of magma containing soda ⇄ Products of dissociation of these silicates in solution in volatiles (mainly water)

Complex silicates of magma containing lime ⇄ Products of dissociation of these silicates in solution in volatiles (mainly water)

2. The fixing of soda compounds in the xenolith results in this:

Complex silicates of magma containing soda → Products of dissociation → Soda being fixed in solid phases in the xenolith

3. At the same time the lime compounds liberated from the xenolith produce this:

Lime compounds liberated from xenolith → Products of dissociation of complex lime silicates of magma in solution in the volatiles → Complex lime silicate molecules of magma

If only reaction (2) took place there would be a piling-up of the products of dissociation other than the soda compound which is removed. The presence of the other reaction (3) prevents this piling-up in excess of the solubility of the various products, because they are utilised in the formation of the complex lime silicates. Thus reciprocal reaction, according to this view, is a necessity in any case where material is added in this way to a xenolith by its enclosing magma.

It will have been noticed that, during this process, the volatiles which enter the xenolith are not actively engaged except in as far as they are required to enter into the composition of the new solid phases being formed in the xenolith. Their main, and very important, function during reciprocal reaction is to act as an intermediary between the complex silicate molecules of the magma and the xenolith. *They form a medium of low viscosity in which the dissociation*

products of the complex silicates and the compounds liberated from the xenolith can diffuse with comparative freedom.

Reciprocal reaction is one way in which a magma gains the material which leads to its contamination. So far, we have dealt only with the behaviour of the xenoliths or contact rocks during reciprocal reaction. Now we must direct some attention to what happens in the magma itself.

In dealing with the elements which enter the magma during reciprocal reaction, there is not the same amount of evidence as to their behaviour as there is in the case of the xenoliths. One of the chief reasons for this is the material gained from the xenoliths by later mechanical disintegration which, when an analysis is made, will tend to mask the changes which can be ascribed to reciprocal reaction alone. The actual crystallisation of new phases, not present in the normal rock, shows, however, that the magma may be considerably affected by the process. Often, too, the effect of reciprocal reaction in the magma is to be seen in the loss of various constituents which are found in increased quantity in the xenolith. We find that when potash is being lost by the magma (Mg,Fe)O tends to be gained, and when soda is being lost, lime tends to be gained. Almost any oxide may be gained by the magma, however, in various examples. One of the reasons for this more elastic behaviour is not far to seek. Acid magmas, though varying in composition, can provide only a limited set of elements, i.e., those already named as being introduced to the xenoliths. Xenoliths, on the other hand, vary widely in composition in different examples and have no such limitations imposed upon them. Also, in the xenolith, as we have already seen, the object of the entry of material is to stabilise the mineral assemblage. The elements entering the magma, on the other hand, are only concerned in giving that magma some composition between that of the xenolith and that of the original magma. And here it should be noted that the xenolith is stabilised with respect to the contaminated magma and not with respect to the original magma. This is a point which Bowen fails to make clear in his discussion of the subject.[46] Reciprocal reaction enables phases to crystallise from the contaminated magma which could not have crystallised from the normal

[46] *Op. cit.*, pp. 197–99, 215–19.

magma, and it is with this modified magma that the xenolith is concerned. The whole object of reciprocal reaction, as we see it, is to arrange things so that both the xenolith and the contaminated magma are in phasal equilibrium without regard to the phases present in either the parent-rock or in the original magma, and if reciprocal reaction can go to completion, then all the phases represented in the xenolith will be represented in the contaminated magma and perfectly stable in it if, and when, they are set free from the xenolith by later mechanical disintegration.

The elements gained by the magma during reciprocal reaction will actually be incorporated in phases which crystallise out from the magma. Consequently the amount and also the nature of the solid phases which crystallise may be altered. Bowen, as is now well known, has dealt in some detail with aspects of reaction between xenolith and magma. Regarding a granite magma as normally saturated, he points out that the magma is not free to become impoverished in constituents at random and that the material used to effect changes in the xenoliths cannot be regarded as simply subtracted from the liquid. He says, "In general, impoverishment in any substance will cause the liquid to pass within a region of saturation and induce the precipitation of some of the phases with which the liquid is saturated."[47] On a later page[48] he cites as an example the subtraction of salt from a solution at $-20°$ C. and says, quite rightly, that ice, not water, will be left as the result. According to our view, however, this is not at all a strict analogy. In reciprocal reaction, the salt would be removed, but another substance would enter in to replace it, so that the formation of ice would not be a necessary consequence. It might or might not be precipitated. Bowen assumes all through that material is being subtracted from the magma, but pays little attention to the material which must be entering the magma from the xenoliths. The fact that all the elements given to the xenolith may reach proportions higher than those found in the crystallised invading rock is strong evidence for assuming that minerals in which those elements would normally be present were not crystallising at the time. Reciprocal reaction would, in general, seem to take place before crystallisation in the magma has attained any sig-

[47] *Ibid.*, p. 197. [48] *Ibid.*, p. 217.

nificance. The question is not only one of heterogeneous equilibria, but also one of homogeneous equilibria. The elements entering the magma from the xenoliths will shift the equilibria already established in solution, but this may or may not lead to the immediate precipitation of solid phases or to the precipitation of new phases.

With regard to the formation of new phases we may study the effect which an influx of lime would have on a normal biotite granite magma, lime being one of the commonest materials introduced by reciprocal reaction. Suppose a quantity of lime to be introduced. If all this lime can be bound in the anorthite molecule in the magma and there is still alumina in excess, then the (Mg,Fe)O of the magma enters into biotite, which is formed at the expense of the potash-felspar molecule. If, on the other hand, there is still lime in excess after the anorthite molecule has been satisfied, then it will combine with some or all (depending on the amount of excess lime) of the (Mg,Fe)O of the magma to form hornblende. In this case the potash which would have combined to form biotite goes to form felspar instead. In this manner there may be a fairly basic plagioclase formed together with biotite and a decrease in the potash-felspar content; there may be a less basic plagioclase and hornblende formed, leaving the potash felspar in greater amount; or there may be plagioclase, hornblende, biotite, and potash felspar, all depending on the amount of lime introduced and the excess left over after the anorthite molecule has been satisfied. Quartz will, of course, be present in greater or less quantity in all the foregoing assemblages, but it occupies a more or less passive rôle, being absorbed in some reactions, thrown out in others. If, during these reactions, potash was being introduced into the xenoliths or (Mg,Fe)O was being introduced into the magma or both were taking place simultaneously, then the potash felspar might almost entirely disappear from among the crystallised products of the magma.

With greater quantities of lime entering the magma, a pyroxene of the diopside-hedenbergite series is formed. In nearly all these examples where hornblende and pyroxene are found, the titanium originally present in the biotite molecule forms sphene or else ilmen-

ite.[49] Thus the effect of lime on an ordinary granite magma is to produce rocks of tonalitic, granodioritic, dioritic, or even ultrabasic nature,[50] as deduced by Bowen.[51]

In the case of pegmatites and aplites, similar relations hold. We have hornblende-bearing pegmatites, the hornblende and its accompanying sphene being formed from the biotite molecule by reaction with the lime introduced.[52] Following this we have the diopside pegmatites[53] and aplites,[54] again usually carrying sphene in some abundance. A pegmatite or aplite, however, usually represents a comparatively small bulk of magma compared with the ordinary granite intrusive. Consequently reaction may go farther here and lead to the formation of phases which would be unstable in any larger mass. We may call these "pseudo-stable phases." Thus in the dissogenites such minerals as grossular and wollastonite may crystallise out from the modified magma.[55]

With soda granites the same thing happens and the assimilation of lime produces melanocratic syenites of one sort or another.[56] Nepheline syenites also become more basic through assimilation of lime.[57]

Summing up, the absorption of lime in acid magmas produces

[49] G. D. Osborne, "The Contact Metamorphism and Related Phenomena in the Neighbourhood of Marulan, New South Wales," Part I, *Geol. Mag.*, Vol. LXVIII (1931), pp. 312–13.

[50] A. Lacroix, "Le Granit des Pyrénées," *Bull. Serve. Carte. Geol. France, No. 71* (1900); B. K. Emerson, "Geology of Massachusetts and Rhode Island," *Bull. U.S. Geol. Surv., No. 597* (1917), pp. 84, 209–10; A. K. Coomaraswamy, "Observations on the Tiree Marble, with Notes on Others from Iona," *Quar. Jour. Geol. Soc.*, Vol. XLIX (1903), p. 96.

[51] *Op. cit.*, p. 218.

[52] A. Laitakari, "Le Gisement de Calcaire Cristallin," *Bull. Comm. Geol. Finlande, No. 46* (1916), p. 23.

[53] "Pegmatitgänge, die Kalkassimiliert haben," *ibid., No. 54* (1920), p. 7; P. Eskola, "On Contact Phenomena between Gneiss and Limestone in Massachusetts," *Jour. Geol.*, Vol. XXX (1922), pp. 291–93; Read, "A Diopside-bearing Pegmatite Near Ellon in Aberdeenshire," *Trans. Geol. Soc. Edinburgh*, Vol. XI (1925), pp. 353–56; E. H. Watson, "A Diopside-bearing Pegmatite in Dolomite," *Econ. Geol.*, Vol. XXIV (1929), pp. 611–25.

[54] Emerson, *op. cit.*, pp. 19, 152.

[55] Lacroix, *Mineralogie de Madagascar*, Vol. II (1922), pp. 373–76.

[56] *Ibid.*, pp. 585–89.

[57] *Ibid.*, pp. 659–61.

more basic rocks in every case. There seems to be little evidence for the view that this absorption can desilicate a granite magma to such an extent that it can produce nepheline syenite as a residue, even if silica is being transferred to the xenoliths or contact. Where nepheline syenites do occur it appears that the magma was already alkaline and poor in silica. We have seen that transference of soda into limestone xenoliths and of potash into dolomitic xenoliths is a process which commonly occurs in granitic magmas, the magma receiving lime and magnesia in exchange. It is difficult to see how this is going to allow a portion of the remaining magma to become rich enough in alkalies to produce a nepheline syenite, but it is quite easy to see how it may give rise to more basic rocks.

A very good example of a reverse process to that outlined above has been observed by Jung.[58] In this instance, a pegmatite already carried some lime in its plagioclase felspar. It has absorbed magnesia by reciprocal reaction with the enclosing rocks and this, combining with the lime from the original felspar, has produced both hornblende and diopside in the pegmatite. It is perhaps worth noting that in alkaline-acid magmas, such as those of nepheline syenite, the absorption of lime almost invariably causes the crystallisation of melanite garnet. This must be regarded as a stable mineral under these conditions, and its formation is probably due to the fact that these magmas are usually comparatively rich in alumina.

A succession of stages can be similarly defined when alumina is entering the magma. Here the genesis of cordierite seems to be as far as any large body of acid magma, gaining alumina with, or more rarely without, magnesia and iron, proceeds. With small bodies such as pegmatite and aplite veins, the change may go yet farther, producing spinel or corundum, but then conditions are rather different.

If corundum or spinel are to be produced, the magma must be desilicated to a certain extent by reaction with its containing walls or xenoliths. This is the case, for instance, in corundum aplites and pegmatites.[59] Such rocks as nepheline syenites, already rich in

[58] J. Jung, *Mem. Serve. Carte. Geol. Alsace et Lorraine, No. 2* (1928), pp. 91–93.

[59] Lacroix, *op. cit.*, pp. 457–63; A. L. du Toit, "Plumasite (Corundum-Aplite) and Titaniferous Magnetite Rocks from Natal," *Trans. Geol. Soc. S. Africa*, Vol. XXI (1918), pp. 53–62.

alumina and comparatively poor in silica, require little desilication to produce corundum or spinel. In these corundum-bearing rocks it is usual to find that potash felspar is absent,[60] so that muscovite can no longer form by combination of the alumina with the potash-felspar molecule. It is curious to find that paragonite does not form, especially as it is a mineral rich in alumina. Apparently the sodic molecules are held so firmly in other combinations that the reaction is inhibited. Although for one corundum syenite, at least, a primary origin for the corundum is postulated,[61] yet for most of these occurrences, and especially for those where the original rock was of granitic composition, the presence of corundum must be due to reaction between the magma and its walls or xenoliths.[62] Corundum and spinel in these cases are, again, pseudo-stable minerals—minerals which would not be present in a larger body of magma.

In some ways this genesis of corundum in small bodies of magma is the reverse process to the absorption of alumina by larger intrusions. The occurrence depends on an actual loss of silica from the intrusion rather than a gain in alumina. Corundum-bearing acid rocks are almost invariably found in places where they cannot have gained any appreciable amount of alumina. A typical example is that described by Du Toit.[63] The normal aplite contains both orthoclase and quartz, but, where bordered by serpentine, it loses both these minerals and is then composed of oligoclase and corundum. The potash of the orthoclase molecule goes into the surrounding serpentine to form an inner zone of phlogopite rock and the silica forms an outer zone of talc rock. This leads to another point in connection with these rocks. At first sight there appears to be no evidence of reciprocal reaction and yet, in the ordinary course of events, material could not be added to the surrounding rocks without causing expansion and disruption, unless there was some reciprocal inter-

[60] A. E. Rogers, "On Corundum-Syenite (uralose) from Montana," *Jour. Geol.*, Vol. XIX (1911), pp. 748–51.

[61] T. H. Holland, "The Swamalai Series of Elaeolite-Syenites and Corundum Syenite in the Coimbatore District, Madras Presidency," *Mem. Geol. Surv. India*, Vol. XXX (1891), p. 212.

[62] Compare S. J. Shand, *Eruptive Rocks* (London, 1927), p. 242, but see E. S. Larsen, *Econ. Geol.*, Vol. XXIII (1928), pp. 398–433.

[63] *Op. cit.*, pp. 53–62.

change. The reason for the lack of reciprocal reaction here is to be found in the nature of the material displaced from the serpentine. In this case it is water, a volatile substance which can be driven off and expelled, thus allowing the reaction to take place in one direction only.

A similar state of affairs may exist when silica is transferred to a fairly pure limestone or dolomite contact or xenolith. Here the silica may replace carbon dioxide which, being a volatile substance, is again driven off and allows the reaction to proceed in one direction only. It is conceivable that, under these conditions, the presence of the limestone might desilicate the magma so as to give a syenitic product from original granitic magma, but this would happen only under these conditions and if the silica was unaccompanied by alkalies during its transfusion to the limestone.

The great influence of reciprocal reaction both on the xenoliths and on the magma is sometimes well brought out by the way in which the modal composition of the contaminated rock changes when its xenoliths also change in composition, and by the way in which the minerals of a xenolith will change to keep pace with changes in the contaminated magma. It is remarkably illustrated, as far as the magma is concerned, by some pegmatites described by Ball.[64] In this case it is not xenoliths, but the country rock immediately surrounding the pegmatites, which has been the cause of change. These pegmatites consist essentially of quartz, alkali felspar, and magnetite. Biotite is characteristic of them when they cut granite, quartz monzonite gneiss, quartz diorite, and hornblende gneiss, though in many places hornblende occurs in pegmatites cutting the last two formations. The pegmatites cutting quartz monzonite areas are typically without ferromagnesian minerals, but contain a soda-lime felspar instead of an alkali one. They also contain allanite, which is an accessory in the quartz monzonite itself. Finally, where the pegmatites cut the Idaho Spring formation (biotite-sillimanite gneiss), muscovite is present. The influence of the country rock on the mineral composition of pegmatite has also been noticed in Coll.[65]

[64] S. H. Ball, "Economic Geology of the Georgetown Quadrangle," *U.S. Geol. Surv. Prof. Paper 63* (1908), p. 64.

[65] "Geology of Ardnamurchan, N.W. Mull, and Coll," *op. cit.*, p. 13.

Summing up, we may say that the object of reciprocal reaction is to produce, as far as possible, the same mineral phases in both the xenolith and the contaminated magma. This, of course, does not necessarily mean that all the phases found in the contaminated rock will be present in the xenolith, but such phases as are present in the xenolith will also be present in the contaminated rock. An exception to this is found in certain occurrences, such as the corundum-bearing rocks mentioned above, where the material displaced from the contact rock is such that it may be driven off and allow reaction to proceed in one direction only. In the ordinary way there is a strong tendency for the minerals thus formed to be of the normal igneous rock-forming series. The occurrence of abnormal minerals is, in general, due to one or the other of two causes. In the xenoliths, unstable minerals may remain, owing to lack of time for reaction. These may also be found in the contaminated rock when the xenoliths are disintegrated. They are to be regarded as unstable relics. In the contaminated magma itself, abnormal minerals may crystallise out as a result of reciprocal reaction. The formation of these minerals is due to the fact that the body of magma in which they are formed is small and so can be more deeply affected. They are the pseudo-stable minerals. Xenoliths associated with such a magma would also carry abnormal minerals due, in this instance, to lack of sufficient material to effect their alteration and not to lack of time.

It is a noteworthy fact that silica appears to play a more or less passive rôle during reciprocal reaction. We have seen that the oxides soda, potash, alumina, and titania may reach a higher figure in the xenolith than in either the invaded or the invading rock (as seen in the solid state). In all the examples yet described, however, the percentage of silica in the xenolith invariably lies at some intermediate value.

It is perhaps necessary to emphasise that, during reciprocal reaction, the magma does not provide material in the form of definite mineral species which are then found in the xenolith. This introduction of mineral species such as quartz and felspar is quite distinct from, and later than, the true period of reciprocal reaction. It is, in fact, the first stage in the disintegration of the xenoliths by the magma and is thus dealt with under the next heading.

MECHANICAL DISINTEGRATION

The actual gaining of various oxides by reciprocal reaction is not the only way in which the magma may alter its constitution. There are what may be termed mechanical methods by means of which the magma incorporates solid material from the invaded rock usually after reciprocal reaction has, at least partially, stabilised its mineral assemblage as far as possible.

Three main methods of incorporation may be noticed, all of which tend to disintegrate the xenolith or rock at the immediate contact so that, in the end, the individual minerals are set free to take their place in the contaminated magma. The first occurs when the invaded rocks or xenoliths are not in equilibrium with the magma as regards all their phases; the second when they do remain more or less in equilibrium; and the third when magma material is injected *en masse* into the xenoliths or country rocks at the contact.

Type I.—An inclusion may be in equilibrium when altered at an early stage in the cooling history of a magma, but certain of its solid phases may later be out of equilibrium and, if they are low in the reaction series, may be dissolved by the magma. Or the xenolith may enter the magma already possessing phases low in the reaction series. This dissolution of certain phases by the magma will thus disintegrate the xenolith and set free the solid phases which are in equilibrium or which are high in the reaction series to be strewn about in the magma.

An instance of this is provided by the Dhoon (Isle of Man) granite.[66] There the biotite, zoisite, and sphene in the xenoliths were in equilibrium throughout, but the albite and paragonite were not. The magma, in this case, disintegrated the xenoliths by dissolving out the albite and paragonite, thus setting free the other minerals. The focus of attack here was mainly marginal, the magma gradually eating its way in and dissolving out the salic minerals until all but the stable phases had disappeared. These are now found strewn about in the granite, but there is no trace of the albite and paragonite which originally accompanied them.

The incorporation of biotite schist may show very similar phe-

[66] Nockolds, *op. cit.*, "The Dhoon (Isle of Man) Granite," *Min. Mag.*, Vol. XXII (1931), pp. 494–509.

nomena. The biotite and such minerals as garnet, etc. (if present), are strewn about in the contaminated rock, but the felspar and quartz which accompanied them in the xenoliths seem to have vanished completely. It may be suggested that in many of these cases it is a question of large crystals growing at the expense of smaller ones. The biotite flakes produced usually appear to be of much the same size as those normally found in the magma. Indeed, it is a common experience to find that the ferromagnesian minerals of xenoliths grow rapidly to a large size when the xenolith is being altered. The grain size of the quartz and felspar is, however, usually much smaller. During crystallisation of the magma, therefore, the normal-sized salic minerals which are crystallising out will gain some of their material at the expense of these smaller grains, thus dissolving them and so incidentally disrupting the xenolith. This type of solution will only apply if the salic minerals have the same composition as those which are crystallising from the contaminated magma and are of smaller-grain size. If the salic minerals of the xenolith have the same composition, but are of the same size as those crystallising from the magma, then this solution will not take place and they will be incorporated by one of the methods mentioned below.[67]

Thus there are two methods by which solution may take place here. If the salic minerals have the same composition as those in the magma, but are of finer grain, then the salic minerals in the magma may grow at their expense. If, on the other hand, the salic minerals in the xenolith are lower in the reaction series than those of the magma, then they will be dissolved by the magma irrespective of the phases which are crystallising. In both cases the process is endothermic[68] and heat will have to be supplied from some source such as crystallisation. This type of disintegration is therefore not entirely mechanical as it depends partly on solution effects for its accomplishment.

Type II.—In the second type of disintegration and incorporation of material all the solid phases of the xenolith are such that all of

[67] Since this paper was written an extremely fine case of this Type I disintegration taking place at a granite-shale contact has been described by Brammall and Harwood, "The Dartmoor Granites; Their Genetic Relationships," *Quar. Jour. Geol. Soc.*, Vol. LXXXVII (1932), pp. 194–98, where it has been described in detail.

[68] Bowen, *op. cit.*, p. 187.

them are incorporated into the magma in the solid state. In this case minerals from the magma gradually wedge their way into the xenolith or are deposited throughout the xenolith, thus forcing the component minerals apart. The mechanism here is the force exerted by the deposited material while crystallising and growing. This involves an actual expansion of the xenolith which can quite easily take place owing to its lying in a plastic medium. Where the action takes place along a line of contact the same thing holds good, because the expansion due to the wedging-in of magma minerals is taken up by the displacement of the solid phases of the contact rock in the direction of the magma.

This type of disintegration is well seen in the Bibette Head (Alderney) granite, where it has been dealt with in detail.[69] There is first a development of "phenocrysts" of plagioclase felspar in the xenoliths and then, at a later stage, the forcing-in of microperthite and quartz which not only occur marginally, but all through the xenolith. This type starts by being marginal, but later affects the whole mass of the xenolith.

It is not known definitely yet whether the development of porphyritic crystals in the xenolith is a particular feature of this type or not. Some references to similar developments have been given in the paper cited above, and another case has been described where orthoclase phenocrysts have been developed in a basic rock invaded by nepheline syenite.[70] This type of disintegration also appears to have been active in a case where diorite has been intruded by biotite granite.[71] The formation of the "phenocrysts" as an early stage of this type is interesting in that, at least in the Bibette Head occurrence, part of the material was provided by the magma and part by the xenolith. It thus forms a natural transition from reciprocal reaction to mechanical disintegration. In other examples, however, the magma appears to have provided almost the whole of the material.

[69] Nockolds, "The Contaminated Granite of Bibette Head, Alderney," *op. cit.*, pp. 441–48.

[70] T. L. Walker and A. L. Parsons, "The Contact Phenomena of the Nepheline Syenites of Port Coldwell, Ontario," *Univ. Toronto Studies, Geol. Ser. No. 24* (1927), pp. 28–32.

[71] M. Billings, "Petrology of the North Conway Quadrangle," *Proc. Amer. Acad. Arts and Sci.*, Vol. LXIII, No. 3 (1928), p. 103.

This type is quite distinct from the type dealt with below in that there is no injection of acid material *en masse*. In the first place, only the salic constituents of the magma take part, and, in the second place, these salic constituents do not enter the xenolith together or in the proportions in which they are found in the magma, but are quite independent of each other. Where this introduction of magma minerals has taken place to any great extent, it may be a most effective method of disintegrating xenoliths. It has an advantage over Type I in that it is purely mechanical and does not depend on solution effects.

Type III.—In the third type of mechanical disintegration there is actual injection of granitic material *en masse*. This occurs chiefly in rocks which have a well-defined schistosity or many close-set planes of weakness giving rise, both at contacts and in xenoliths, to *lit-par-lit* injection.

In general, we should expect schistose rocks to be disintegrated for the most part by Type III, and others would tend to be disintegrated by Type I or II. Actually this seldom appears to be the whole story. More schistose rocks have often been affected by the earlier stages of Type II before the oncoming of Type III. For instance, Erdmannsdorffer[72] describes the gradual development of porphyroblastic felspars in mica schists with further introduction of granitic material leading to the formation of gneisses and then, at the immediate contact, Type III method comes in giving *lit-par-lit* injection. An exactly similar sequence of events is now known from injection complexes in many parts of the world. Thus here both Types II and III have helped in the disintegration of the rocks at the margin of the intrusive mass. The final stage is the production of one type of foliated granite.[73]

Often, again, Types I and III are associated, as is well shown in some interesting rocks described by Wagner.[74] There a series of schistose biotitic rocks has been permeated and injected by granitic

[72] O. H. Erdmannsdorffer, "Petrographische Untersuchungen an einigen Granit-Schieferkontakten der Pyrenäen," *Neue Jahr. für Min., etc. B.B. 37* (1914), pp. 739–65.

[73] G. A. J. Cole, "On Composite Gneisses in Boylagh, W. Donegal," *Proc. Roy. Irish Acad.*, Vol. XXIV, Sec. B (1902), pp. 203–30.

[74] P. A. Wagner, "Interesting Migmatic Rocks from the Gordonia District, Cape Province," *Trans. Geol. Soc. S. Africa*, Vol. XX (1918), pp. 48–50.

material. All gradations are exhibited from schists veined with granite through migmatitic gneisses to granites in which dark streaks and clots of biotite and sillimanite and isolated porphyroblasts of sillimanite and garnet are all that remain of the original sediments.

More rarely, injection of magmatic material *en masse* takes place in rocks which show no well-defined schistose structure. Some examples of this have been given in some detail from Ardnamurchan.[75] A granophyric magma has been injected *en masse* in basic igneous rocks, so that the acid magma retains its entity and has not introduced its minerals separately as is the case with the Type II method.

Thus, although we have described each type separately, it will be seen that, in schistose rocks at least, two of the types may be associated together. Types I and II are, of course, incompatible and cannot both be found together in the same rock. When we are dealing with rocks which have igneous or hornfels textures, Type III does not usually occur, and disintegration is then effected by Type I or II. Even in these there is considerable difference of behaviour, but very little is known about this side of the question at present. We should expect that a sutured type of grain structure would be very resistant to disintegration. The equigranular texture exhibited by some hornfelses and metamorphosed limestones should also be difficult to disintegrate unless certain phases were low in the reaction series when solution effects might intervene. Igneous rocks (plutonic or hypabyssal), on the other hand, should not be so difficult. This is especially the case with dolerite, for instance, where the lath-shaped felspars produce straight-edged boundaries which are potential planes of weakness and along which the introduced magma minerals can work their way. The texture does not, of course, affect Type I as much as Type II because the former, as already stated, depends on solution effects. These observations do seem to be borne out by such facts as are at present to hand. Between the easy disintegration of many schistose rocks and the resistance of rocks with strong interlocking and sutured textures there will be all gradations, and in any large rock series which is broken up by, or comes in contact with, a magma each rock type will react to mechanical disintegration on its own merits.

[75] "Geology of Ardnamurchan, N.W. Mull, and Coll," *op. cit.*, pp. 276–77; 278–79.

Movement in the magma is a potent aid to mechanical disintegration. It tends to distribute the solid phases set free from the xenolith in a uniform manner throughout the contaminated magma, thus obscuring its hybrid character. It must also favour the introduction of magmatic minerals into the xenoliths by helping to provide fresh supplies of material.

The products given to the magma by mechanical disintegration are either small clots or else individual minerals. These minerals are, strictly speaking, xenocrysts. The term "xenocryst" is usually given to a crystal which has been derived from some outside source and *is foreign to the igneous rock in which it lies*. If reciprocal reaction has been able to go to completion, under favourable circumstances and before mechanical disintegration sets in, the minerals set free from the xenolith will certainly have come from some outside source, but they will not be foreign to the rock in which they lie. They will, in fact, be representative of the normal rock-forming minerals of the magma and be just as much at home in the contaminated magma as will the minerals which crystallise out. Thus although these minerals which are derived from the xenoliths must be regarded as xenocrysts, they will by no means always fulfil the second part of the definition given above. Under these circumstances, they may be quite indistinguishable, as such, from the minerals which have crystallised from the magma. In other cases some inherent peculiarity may enable them to be distinguished even when they are of the same species as those crystallising.

CONCLUDING REMARKS

In the present paper we have confined ourselves to a study of how contamination is effected. Lack of space precludes the discussion of such topics as the outside factors which must influence the contamination processes. Further, we have said nothing about the ability of these processes to produce normal rock types. In the writer's opinion such normal types can be produced under certain favourable conditions.[76]

[76] The writer wishes to express his best thanks to Professor C. E. Trilley, of the Department of Mineralogy and Petrology, Cambridge; to Professor O. T. Jones of the Sedgwick Museum, Cambridge; and to Dr. A. Harker, Emeritus Reader of Petrology in the University of Cambridge, for much helpful discussion and advice on the various problems dealt with in this paper.

BOULDERS FROM BENGALIA

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ABSTRACT

This paper treats of the boulders in the Carboniferous Caney shale in the Ouachita Mountains of southeastern Oklahoma. Some of the boulders are composed of local rock, but others are composed of rock which does not crop out in the Ouachitas. As theories of long-distance transportation of the so-called exotics are considered inadequate, a theory is developed which explains the origin of the boulders from a local source.

Evidence is cited which indicates that rocks like those which comprise the apparently foreign boulders underlie certain areas in the Oklahoma Ouachitas. A land, Bengalia, existed along the north side of the Ouachita area throughout most of Paleozoic time. Crustal warpings caused variations in the positions of land and sea areas, and erosion exposed the older rocks. During most of Mississippian time, Bengalia formed the north barrier of the filter-basin-like Ouachita trough in which the thick Stanley and Jackfork formations were deposited while only a few hundred feet of lower Caney shale were being laid down to the north and west. By later Caney time, Bengalia and Llanoria, the land to the south, were both lower, and straits connected the trough with seas to the north and west. Earth tremors, accompanying downwarping of the trough, caused submarine landslips of boulders from Bengalia into the upper Caney shale then being deposited in the northwest side of the trough.

INTRODUCTION

The problem of the boulders in the Caney shale and adjacent formations of the Ouachita Mountains in Oklahoma has been studied for more than a quarter of a century, and because of its spectacular and baffling nature it has attracted many geologists.

The "Caney" boulders are composed of sedimentary rocks older than the matrices in the respective horizons. As many of them are of rock which does not crop out in the Ouachitas, they have been called exotic. Some of the boulders or blocks are so immense that their transportation from an extra-Ouachita source is doubtful.

The summer of 1930 was spent in the Oklahoma Ouachitas studying the geology of the region with special reference to the boulder problem. Most of the known outcrops of boulder horizons were visited.[1]

[1] The writer wishes to express his appreciation for hospitality shown him by many residents of the region. Thanks are especially due to Mr. Lee Basket of Johns Valley, Mr. Bill Epperly of Bengal, and Mr. Alfred L. Allen of Ti, for helpful guidance in areas near their homes. Mr. H. D. Miser gave several much appreciated suggestions regarding field study. The writer is especially indebted to Professor R. T. Chamberlin and to Dr. Carey Croneis of the Department of Geology at the University of Chicago for many helpful suggestions and criticisms.

PREVIOUS INVESTIGATIONS

The first-discovered "Caney" boulders were thought to be remnants of a limestone ledge, and Taff[2] mapped the underlying Jackfork sandstone and Stanley shale as Ordovician, because fossils of that age were found in some of the blocks. Evidence of the post-Ordovician age of these formations, however, was soon obtained.[3] Further study of the occurrence of the boulders showed that they were neither intra-formational nor basal conglomerate fragments. Taff[4] briefly described the boulder horizons, and Woodworth[5] described the outcrop at Compton cut near Talihina, and concluded that scratches found on boulders at that locality are of tectonic rather than of glacial origin. More recently Miser,[6] Ulrich,[7] Powers,[8] and van der Gracht[9] have given more complete accounts of the age, occurrence, and possible origin of the boulders.

STRATIGRAPHY OF THE OUACHITAS

The rocks exposed in the Ouachita Mountains are almost entirely non-limestone sedimentary[10] rocks of Paleozoic age. All systems from Cambrian to Pennsylvanian are at least partially represented. The total thickness of strata is about 25,000 feet of which 15,000–20,000 feet are of Carboniferous age.[11] Sandstone and shale forma-

[2] J. A. Taff, *U.S. Geol. Surv., Geol. Atlas, Atoka Folio, No. 79* (1902).

[3] E. O. Ulrich, "Fossiliferous Boulders in the Ouachita Caney Shale and the Age of the Shale Containing Them," *Okla. Geol. Surv. Bull. 45* (1927), p. 5.

[4] "Ice-borne Boulder Deposits in Mid-Carboniferous Marine Shales," *Geol. Soc. Amer. Bull.*, Vol. XX (1909), pp. 701–2.

[5] J. B. Woodworth, "Boulder Beds of the Caney Shale at Talihina, Oklahoma," *Geol. Soc. Amer. Bull.*, Vol. XXIII (1912), pp. 457–62.

[6] H. D. Miser, "Structure of the Ouachita Mountains of Oklahoma and Arkansas," *Okla. Geol. Surv. Bull. 50* (1929), pp. 27–30.

[7] *Op. cit.*

[8] Sidney Powers, "Age of the Folding of the Oklahoma Mountains," *Geol. Soc. Amer. Bull.*, Vol. XXXIX (1928), pp. 1031–72.

[9] W. A. J. M. van der Gracht, "The pre-Carboniferous Exotic Boulders in the So-called Caney Shale in the Northwestern Front of the Ouachita Mountains of Oklahoma," *Jour. Geol.*, Vol. XXXIX (1931), pp. 697–715.

[10] Several small outcrops of igneous rocks ranging in age from possible Ordovician to Upper Cretaceous occur in the Arkansas and McCurtain County, Oklahoma, areas.

[11] H. D. Miser and C. W. Honess, "Age Relations of the Carboniferous Rocks of the Ouachita Mountains of Oklahoma and Arkansas," *Okla. Geol. Surv. Bull. 44*, p. 7.

tions of Cambrian and early Ordovician age are confined in outcrop to Arkansas and McCurtain County in the southeast corner of Oklahoma. The remaining rocks of the older Paleozoic and Devonian systems are mostly shales and unusually thick, bedded flints and cherts which have a wide distribution in Arkansas and McCurtain County, Oklahoma. These beds also crop out in the Potato Hills, in the north-central part of the Oklahoma Ouachitas, and in Black Knob Ridge at the west end of the province, where they are known as the Stringtown shale (Ordovician) and Talihina chert (Ordovician to Mississippian?). Arkansas novaculite (Devonian to Mississippian), which is equivalent to upper Talihina chert, also crops out in five other small inliers on the north-northwest side of the Oklahoma Ouachitas. The Blaylock sandstone (Silurian) is confined in outcrop to McCurtain County, Oklahoma, and to Arkansas. Much of Silurian time appears to be represented by hiatuses.

The Hot Springs sandstone which is present in the eastern end of the mountains in Arkansas, and the thick Stanley shale (0–10,000 feet) are of Carboniferous age, and unconformably overlie the older rocks. The thick Jackfork sandstone (0–6,600 feet) conformably overlies the Stanley shale. The Stanley shale and Jackfork sandstone have the most widespread exposures of any formations in the Ouachita Mountains, but they are not present along the north side of the province in Oklahoma.

The Caney black to gray shale is the next younger Carboniferous formation, and is exposed along the north side of the mountains in Oklahoma. Along the northern border of the province the formation lies on the Arkansas novaculite, part of which is the equivalent of the Woodford chert, and the Stanley and Jackfork formations are not present. This stratigraphic relation of the Caney shale to older rocks, which is the same in the Arbuckle Mountains about 15 miles to the west of the Ouachitas, is shown in two small areas along the northwest arc of the mountains. At the Wesley inlier the sequence is Arkansas novaculite, Caney, Wapanucka,[12] and Atoka, but, at the Brushy Creek inlier, which is 8 miles northeast of Wesley, the Wapanucka has not been recognized. At Brushy Creek there are cherty limestone beds which are not the equivalent of any rocks in the Ar-

[12] Powers, *op. cit.*, p. 1042.

buckle Mountain section, but which are the equivalent of the lower division of the Arkansas novaculite, and there are overlying beds which correspond to younger portions of that formation. The absence of the Stanley and Jackfork formations along the north side of the mountains farther to the east is also shown at two localities near Bengal, at which the Atoka formation rests on Arkansas novaculite. Here the Caney and Wapanucka formations, as well as the Stanley and Jackfork, are missing. This relation also appears to exist at another small inlier of novaculite on the northwest arc, 7 miles south of Hartshorne.

Farther south, but in the north half of the Oklahoma Ouachitas and near their west end, a black shale conformably overlies the Jackfork sandstone. The age of this shale is a matter of controversy. Most geologists who have published opinions on the subject believe that this shale is the Caney. In fact, the type locality of the Caney shale which is Johns Valley, formerly called Caney Basin or Cove,[13] is in this region where the black shale overlies Jackfork sandstone. From the lower part of the black shale in Johns Valley, Miser[14] has collected a fauna, all of the fossils of which "belong to the fauna of the Mississippian Caney shale." This fauna was found in phosphate nodules and "concretionary masses of limestone." The criteria for distinguishing between concretionary masses of limestone and exotic blocks and boulders of older rock appear to be (1) that the concretions lie parallel to the bedding, (2) that the lithology of the shale including the character and distribution of nodules and limestone concretions is the same as in the Caney where it rests on Woodford, and (3) that at the localities of the collection all the fossils and rock are of Caney age. Miser believes that the fauna is indigenous, and that the shale which lies on the Jackfork is Caney shale and contains beds of both Mississippian and Pennsylvanian age.[15] Ulrich,[16] however, believes that Miser's fauna is not indigenous, but that it was transported into the shale, and that the shale which lies on the Jackfork has no known fauna or flora of its own. He has proposed the name, "Johns Valley shale," for the black shale overlying Jackfork sandstone, and refers both of these formations to the Lower Penn-

[13] Miser and Honess, *op. cit.*, p. 11.

[14] *Ibid.*, footnotes on pp. 22 and 23.

[15] *Ibid.*, p. 28.

[16] *Op. cit.*, p. 23.

sylvanian, because of the few, not very diagnostic, fossils from the Jackfork, and because a formation of coarse clastic material is such a one as might be expected to be the initial formation of a system.

The Wapanucka limestone overlies the Caney shale along the north border of the Ouachitas in Oklahoma and in the Arbuckle Mountains. This formation contains such a definitely very early Pennsylvanian fauna that, on this evidence alone,[17] it may be thought that the 6,000–7,000 feet of Jackfork sandstone which underlies a sandstone bed containing this fauna in northwestern McCurtain County, Oklahoma,[18] cannot also be of Pennsylvanian age. Wapanucka limestone crops out on Windingstair Mountain[19] near Bengal, several miles south of the north border of the mountains, where the succession is Jackfork, Caney, and Wapanucka. Limestone of probable Wapanucka age was also found on the north slope of Windingstair Mountain at a new locality at Bill Epperly's prospect $2\frac{1}{2}$ miles east of Bengal (S. $\frac{1}{2}$ Sec. 8, T. 4 N., R. 22 E.). Wapanucka strata have not been reported from the interior of the mountains except at the Bengal and McCurtain County localities.

Succeedingly younger Pennsylvanian formations are the Atoka sandstone and shale formation, the Hartshorne sandstone, the McAlester shale, the Savanna sandstone, and the Boggy shale. These formations crop out along the north and northwest border of the Ouachita Mountains, in the Arkansas Valley, and on the east end of the Arbuckles. The Boggy shale is the youngest formation which is definitely known to have been folded with the Ouachita Mountains. Sandstone and shale strata preserved in synclines in the interior of the western end of the Ouachitas overlie Caney shale, and have been referred to the Atoka. Wapanucka beds have not been recognized in these synclines, the easternmost of which is Johns Valley. In the Boktukola syncline in northwestern McCurtain County several thousand feet of sandstone beds overlie the Jackfork sandstone. The sandstone above the Jackfork is thought to be Atoka,[20] because a

[17] Carey Croneis, "Pennsylvanian Overlap," *Amer. Assoc. Petr. Geol. Bull.*, Vol. XV (1931), pp. 471–73.

[18] C. W. Honess, "Geology of Southern Leflore and Northwestern McCurtain Counties, Oklahoma," *Bur. Geol. Cir. 3* (Norman, Okla., 1924).

[19] Powers, *op. cit.*, p. 1043.

[20] Miser and Honess, *op. cit.*, p. 21.

thin sandstone containing a Wapanucka fauna[21] intervenes between the two formations. There is no evidence of interruption in sedimentation, and, accordingly, the Caney formation is probably represented by sandstone strata in the upper part of the Jackfork. The Atoka overlies the Jackfork in two synclines extending east-west across Clark and Pike counties, along the south border of the Ouachitas in Arkansas. Atoka also overlies Jackfork along the north side of the Arkansas Ouachitas and in the Arkansas Valley in Arkansas, the Caney shale being essentially absent. Younger rocks than the Atoka have not been recognized in the interior of the Ouachitas.

DISTRIBUTION OF BOULDERS

Boulders and blocks of rock in Caney shale and adjacent formations are confined in outcrop to the north side and the west end of the Oklahoma Ouachitas. Fragments of limestone in a Wapanucka breccia, described by Wallis[22] from the east end of the Arbuckles, have only pebble dimensions (1 mm.—3 cm.). Most of the boulder horizons are in the southernmost outcrops of "Caney," presumably where the formation lies on Jackfork. At the west end of the mountains there are boulder horizons in Caney on Jackfork. There are, however, three boulder localities in that part of the "Caney" which probably does not overlie Jackfork, but which overlaps the Arkansas novaculite. Two localities are south of Recyl (Sec. 25, T. 4 N., R. 17 E.,[23] and a new locality, Sec. 11, T. 3 N., R. 17 E.), and the third is 8 miles east of Bengal[24] (on section line between Sec. 7 and 8, T. 4 N., R. 23 E.). Boulder horizons crop out from about 10 miles east of Atoka eastward to the vicinity of Boles, Arkansas, about 20 miles east of the Oklahoma border.

Boulder horizons may be grouped into two quite general stratigraphic ranges. The lowest is about 5–10 feet[25] above the base of the Caney or possibly in the top of the Jackfork, as in Johns Valley (below Miser's Mississippian Caney fauna) and near the top of the

[21] *Ibid.*, p. 14.

[22] B. F. Wallis, "The Geology and Economic Value of the Wapanucka Limestone of Oklahoma," *Okla. Geol. Surv. Bull.*, XXIII (1915), p. 72.

[23] Powers, *op. cit.*, p. 1041.

[24] Ulrich, *op. cit.*, p. 11, locality No. 27.

[25] Powers, *op. cit.*, p. 1043.

Jackfork north of Compton cut.[26] The other horizons are in Pennsylvanian rocks in the upper Caney (Springer), in Wapanucka[27] shale, and in the Atoka at Stapp.

Boulders are embedded in black or gray shale matrices, and in a sandstone matrix in the Atoka horizon at Stapp. Because of these matrices the beds do not resemble friction or fault surface breccias. Boulders are subangular and rounded, and many are corroded. They are arranged in beds in the shale at some localities and boulders of several different kinds of rocks commonly occur in the same bed.

ARBUCKLE SECTION COMPARED WITH ROCKS IN THE BOULDERS

The east end of the Arbuckle Mountains lies only a few miles west of the Ouachitas, but the Paleozoic rocks there exposed are very different from those which crop out in the Ouachitas, and yet Arbuckle-like rocks have largely contributed to the Caney boulders. Ulrich[28] has studied faunas found in the boulders, and conclusions respecting their identity are taken from his paper.

Paleozoic rocks overlie a pre-Cambrian basement of Tishomingo granite at the east end of the Arbuckles. Pre-Cambrian and Cambrian rocks are not represented in Ouachita boulders, and in Ulrich's collection only three fossils taken from boulders represent the upper part (Canadian) of the Arbuckle (Cambro-Ordovician) limestone. With the exception of McCurtain County, nowhere in the Oklahoma Ouachitas do rocks of Arbuckle or greater age crop out.

The Simpson-Bromide sandstone, shale, and limestone formation and the younger Viola limestone unconformably succeed the Arbuckle limestone in the Arbuckle Mountains. These formations are of Middle and Upper Ordovician age. Rocks representing this time interval in the Ouachitas are the Stringtown or Womble shale, the Bigfork chert, and the Polk Creek shale. Numerous fossiliferous boulders of Middle and Upper Ordovician age have been found in the Ouachita "Caney," and many of the fossils are identical with those from the Simpson, Bromide, and Viola, but about one-fifth of the species have not been found in place in the Arbuckles, and a Lower Simpson fauna has not been found.[29] A single boulder is referred to

[26] *Ibid.* [27] *Ibid.* [28] Ulrich, *op. cit.*, pp. 12–21. [29] *Ibid.*, p. 16.

the Sylvan shale. Rocks in Ordovician boulders are more like Arbuckle rocks than they are like outcropping Ouachita rocks, but they differ somewhat from the former.

Taff's[30] Hunton formation in the Arbuckles has been divided by Reeds[31] into five formations: the Chimneyhill limestone and Henryhouse shale of Silurian age, and the Haragan shale, Bois d'Arc limestone, and Frisco limestone of Lower Devonian age. Very much of Hunton time appears to be represented by hiatuses in the Ouachita section in Oklahoma which is exposed in the Potato Hills and Black Knob Ridge. The upper part of the Talihina chert in these two inliers includes the Missouri Mountain slate and the lower, middle, and upper Arkansas novaculite. The slate may partially represent the Devonian portion of the Hunton. Oölitic limestone boulders of Chimneyhill (Silurian) age were found at two widely separated localities. Chimneyhill time is represented by a hiatus in the Talihina chert.

Devonian fossils were found in chert blocks from two localities. The rock is distinctly Ouachita rock[32] and is referred to the Arkansas novaculite which crops out in the Brushy Creek inlier 12 miles northwest of the most fossiliferous of the boulders, which was found in Johns Valley. Arkansas novaculite crops out $\frac{1}{2}$ mile and $1\frac{1}{2}$ miles west of Bengal, and the writer was fortunate in finding a new outcrop of the formation (center of N. $\frac{1}{2}$ of N. $\frac{1}{2}$ of Sec. 7, T. 4 N., R. 21 E.) $4\frac{1}{2}$ miles west of Bengal. At the latter locality limestone and shale beds as well as black chert are exposed, but no identifiable fossils were found. This outcrop is within 5 miles to the northeast of Ulrich's other Devonian boulder locality.

Many Mississippian Caney fossils have been found in the Caney shale of the interior Ouachitas. Some of the large shells are silicified and are not in a like matrix. They may have entered the shale as did the boulders, thus being boulders themselves. If Miser's contention, however, that there are Mississippian Caney beds in Johns Valley is correct, these fossils or "boulders" are probably of Ouachita rock.

From three localities, two near Bengal and one near Stapp, Pennsylvanian Wapanucka boulders are reported. Powers[33] has shown

[30] *Op. cit.*

[31] C. A. Reeds, "The Arbuckle Mountains, Oklahoma," *Okla. Geol. Surv. Cir. 14* (1927), p. 12.

[32] Ulrich, *op. cit.*, p. 19.

[33] *Op. cit.*

that there are boulder horizons definitely younger than Wapanucka within a few miles of each of these horizons and doubtless the boulders came from such horizons. Miser[34] and the writer have found outcrops of Wapanucka limestone on Windingstair Mountain near Bengal. In the railroad cut 1 mile south of Stapp a boulder horizon is referred to the Atoka formation,[35] because it has a sandstone matrix. In this sandstone bed, boulders or blocks of black shale containing nodules were found by the writer. The shale fragments were several inches in diameter and undoubtedly are Caney shale.

There are many boulders which contain no fossils, and consequently cannot be definitely placed in the section. Chert and flint boulders are abundant although of relatively small sizes, probably because the rock is friable. All of the Talihina horizons of chert may be represented, including Bigfork chert and Arkansas novaculite. These chert boulders are of Ouachita rocks. There are also many sandstone, quartzite, and shale boulders, but they are neither so conspicuous nor so numerous as limestone and flint boulders.

The thick Stanley-Jackfork sequence is absent in the Arbuckles. This time interval may be represented by a hiatus, or perhaps it is more likely, as suggested by Honess,[36] that Caney shale in the Arbuckles and along the north border of the Ouachitas represents Stanley and Jackfork, as well as the Caney of the interior of the Ouachitas.

In summing up the characteristics of the rocks comprising the boulders, it should be noted that the younger Paleozoic systems are represented by Ouachita rocks which crop out in small areas along the north and northwest border of the mountains. The older Paleozoics are conspicuously represented by Ordovician limestones, and there are a few boulders of Chimneyhill (Silurian) limestone. The immense blocks in Johns Valley are probably of Ordovician age.[37] Boulders of lower Paleozoic rocks are more like the Arbuckle rocks than the outcropping Ouachita rocks, but not nearly all of the fossiliferous limestone formations of the Arbuckles are represented.

[34] In Powers, *op. cit.*, p. 1043.

[35] *Ibid.*, p. 1041.

[36] C. W. Honess in Miser and Honess, *op. cit.*, p. 11.

[37] Powers, *op. cit.*, p. 1043.

THEORIES OF ORIGIN OF THE BOULDER BEDS

Because of the apparently foreign rocks comprising some of the boulders, various theories have been proposed to explain their introduction into the Ouachita "Caney" shale. Long-distance transportation, whether by rafting or by overthrusting, is usually postulated. The Criner Hills[38] area is the nearest known which might have been a land exposing rocks like those in the boulders at the time of their deposition. But the Criner Hills are many miles to the southwest, beyond strata which are of the equivalent age of the boulder horizons, but which contain no boulders.

Floe- or berg-ice rafting appears to be the most generally accepted means of transportation. In Johns Valley, and there alone, immense boulders of more than 30 feet in greatest dimensions are known. Powers[39] and Miser[40] have given measurements of the largest of these great blocks of rock. Miser's dimensions for three blocks are: 200 feet in length, 110 feet by 195 feet, and 50 feet by 369 feet. Each of these three blocks, all found in only the 10 square mile outcrop of Caney in Johns Valley, is twice as large in greatest dimension as any Pleistocene glacial erratic which the writer has found reported from all of North America. Pilot Rock,[41] near Coule City, Washington, one of the largest glacial boulders, has a greatest dimension of less than 100 feet. The Madison boulder in New Hampshire is about 90 feet long and is thought to have traveled about 3 miles.[42]

The large erratics of the Iowan drift have in the past been used as a criterion in identifying that sheet, but the largest boulder reported from Iowa[43] is 50 feet by 40 feet by $11\frac{1}{2}$ feet. There may be no limit to the size of a mass of rock which can be picked up by a sufficiently large glacier, but the crushing and weaving action caused by the variant pressures within the ice during transportation causes an observed decrease in size with distance from the parent ledge.[44] Similar

[38] *Ibid.*, p. 1044.

[39] *Ibid.*, p. 1043.

[40] Miser and Hones, *op. cit.*, p. 22 n.

[41] T. C. Chamberlin and R. D. Salisbury, *Geology*, Vol. III (New York, 1907), Fig. 474.

[42] J. W. Goldthwait, personal communication, May 10, 1931.

[43] G. F. Kay, "Some Large Boulders in the Kansan Drift of Southern Iowa," *Ia. Geol. Surv.*, Vol. XXVII (1930), pp. 345–53.

[44] Goldthwait, *op. cit.*

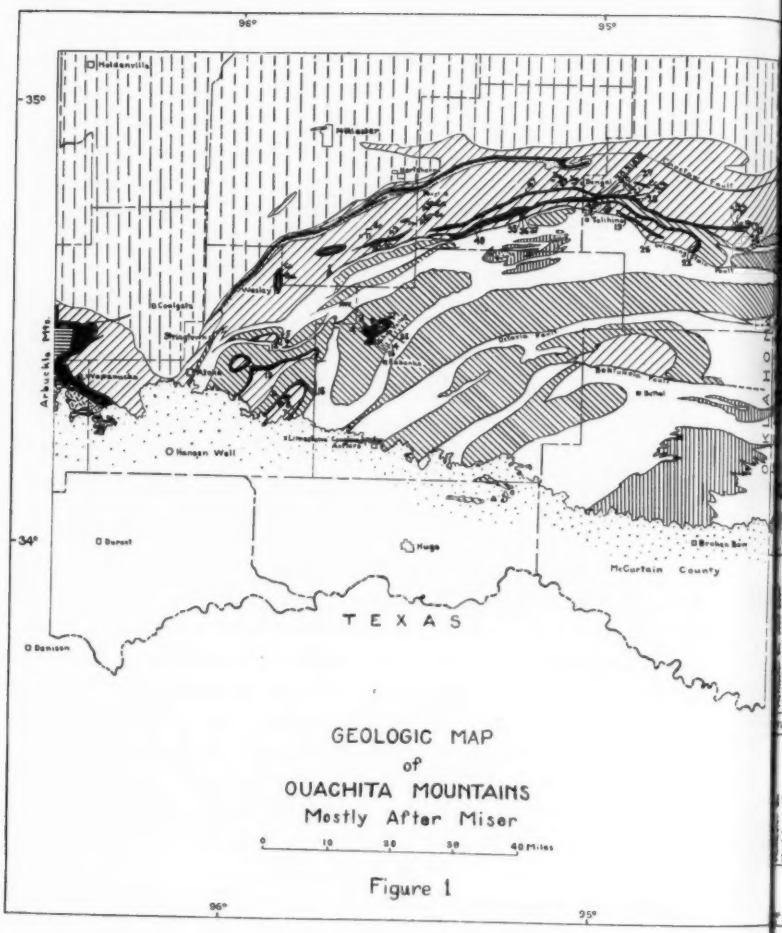
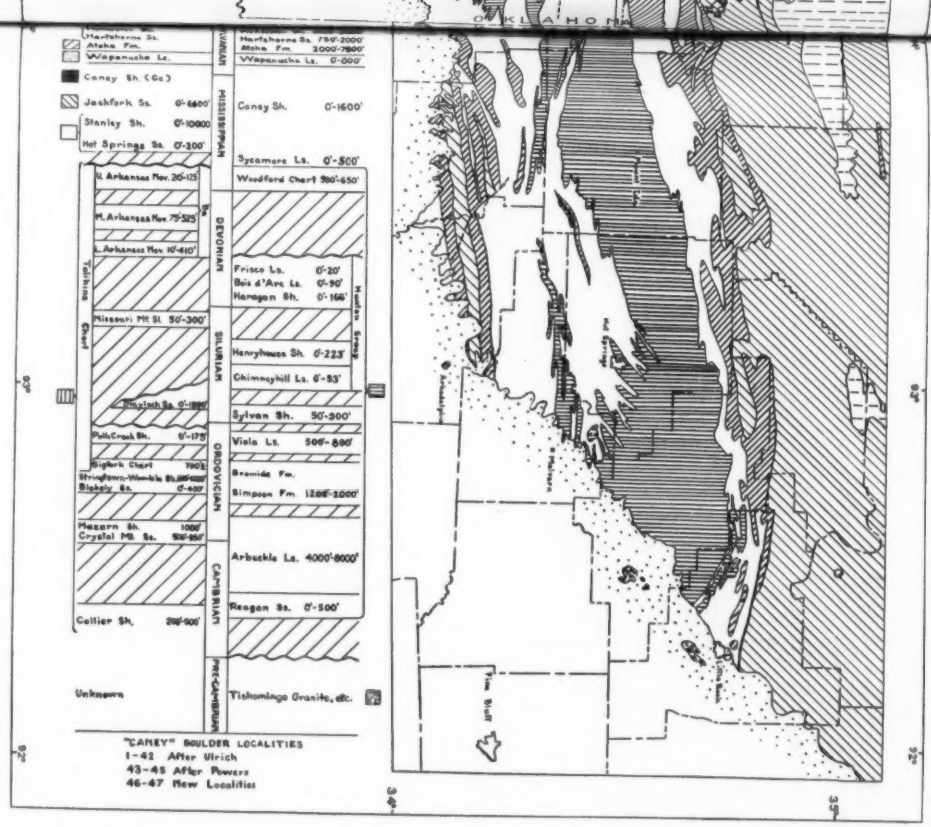

96°
95°
35°
34°
Arbuckle Mts.
Hugo
McCurtain County
TEXAS
GEOLOGIC MAP
of
OUACHITA MOUNTAINS
Mostly After Miser
0 10 20 30 40 Miles

Figure 1


Hartshorne Ss.
Atoka Fm.
Wapanucka Ls.
Caney Sh. (Cc)
Jackfork Ss. 0'-6600'
Stanley Sh. 0'-10000'
Hot Springs Ss. 0'-200'
U. Arkansas Nov. 20-125
M. Arkansas Nov. 75-525
L. Arkansas Nov. 10'-610'
Missouri Mt Sl. 50'-300'
Plattin Ls. 0'-1200'
Plattin Chert
Pole Cr. Sh. 0'-175'
Kingsh Chert 700'
Stringtown-Womble Sh. 0'-600'
Blakely Ss. 0'-620'
Mazarn Sh. 1000'
Crystal Mt. Ss. 200'-250'
Collier Sh. 200'-500'
Unknown
PENNSYLVANIAN
MISSISSIPPIAN
DEVONIAN
SILURIAN
ORDOVICIAN
CAMBRIAN
PRE-CAMBRIAN
Hartshorne Ss. 750'-2000'
Atoka Fm. 2000'-7800'
Wapanucka Ls. 0'-800'
Caney Sh. 0'-1600'
Sycamore Ls. 0'-500'
Woodford Chert 300'-650'
Frisco Ls. 0'-20'
Bois d'Arc Ls. 0'-90'
Haragan Sh. 0'-166'
Hunton Group
Henryhouse Sh. 0'-223'
Chimneyhill Ls. 0'-83'
Sylvan Sh. 50'-300'
Viola Ls. 500'-800'
Bromide Fm.
Simpson Fm. 1200'-3000'
Arbuckle Ls. 4000'-8000'
Reagan Ss. 0'-300'
Tishomingo Granite, etc.
"CANEY" BOULDER LOCALITIES
1-42 After Ulrich
43-45 After Powers
46-47 New Localities
OKLAHOMA
Texas
Red River
Washita R.
Blue R.
Muddy Boggy Cr.
Clear Boggy Cr.
Little Wichita R.
Red R.
Little R.
Kiamichi R.
Mountain Fork R.
Glover R.
Broken Bow Res.
Pine Cr. Res.
Hugo Res.
Sardis Res.
Jackfork Mts.
Winding Stair Mts.
Ouachita Mts.
Arkansas
Fort Smith
Poteau
Talihina
Heavener
Wister
Howe
Spiro
Sallisaw
Stigler
Gore
Muskogee
Tahlequah
Eufaula
Checotah
Henryetta
Okmulgee
Sapulpa
Tulsa
Bartlesville
Nowata
Coffeyville
Independence
Chanute
Parsons
Pittsburg
Miami
Joplin
Baxter Springs
Columbus
Galena
Picher
Commerce
Quapaw
Vinita
Afton
Grove
Jay
Kansas
Missouri
Neosho R.
Verdigris R.
Arkansas R.
Illinois R.
Grand R.
Spring R.
Elk R.
Cowskin R.
Shoal Cr.
Big Cabin
Chouteau
Locust Grove
Salina
Pryor
Claremore
Catoosa
Inola
Broken Arrow
Coweta
Haskell
Porum
Braggs
Webbers Falls
Oktaha
Wagoner
Fort Gibson
Hulbert
Tahlequah
Stilwell
Westville
Kansas
Welling
Collinsville
Owasso
Skiatook
Oologah
Chelsea
Nowata
Bartlesville
Dewey
Ochelata
Copan
Ramona
Caney
Dearing
Sedan
Elgin
Chetopa
Cherryvale
Coffeyville
Independence
Parsons
Oswego
Labette
Neodesha
Altamont
Bolton
Edna
Thayer
Galesburg
Chanute
Erie
Girard
Pittsburg
Frontenac
Arma
Fort Scott
Uniontown
Mapleton
Redfield
Bronson
Hiattville
Mulberry
Pleasanton
Hepler
Moran
Iola
Humboldt

action in a fault breccia under an overthrust sheet must have an even greater cracking effect. Data at hand regarding the maximum size of Pleistocene glacial erratics indicate that it is highly improbable that three such huge blocks could have been transported into the small Johns Valley area by icebergs from glaciers.

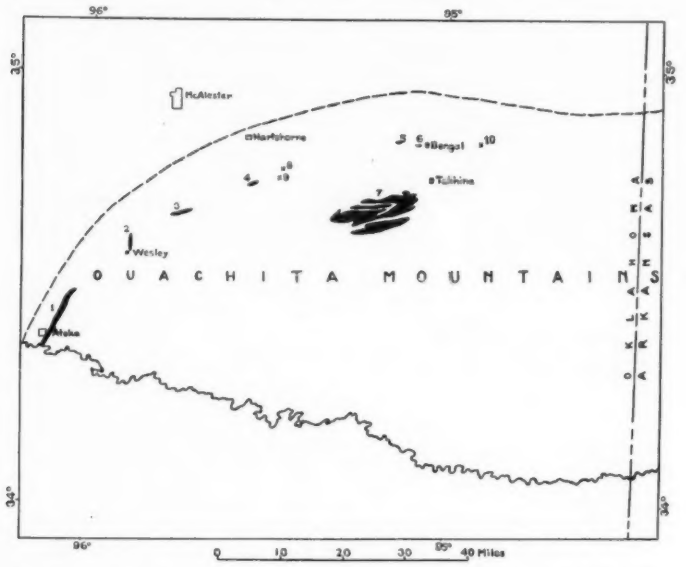


FIG. 2.—Map of southeastern Oklahoma, showing inliers of Bengalia; *1*, Black Knob Ridge; *2*, Wesley; *3*, Brushy Creek; *4*, 7 miles south of Hartshorne; *5–6*, west of Bengal; *7*, Potato Hills, and "Caney" boulder localities along the axis of Bengalia; *8–9*, south of Recyl; *10*, 8 miles east of Bengal.

Kindle[45] has stated that "the ability of ice floes to scour the bottom and transport material in shallow arctic seas when driven by currents or the wind is comparable with that of glaciers." Little is known of the ability of floe ice to pick up and carry large masses of rock; for most of such present-day work is confined to the polar regions and deposits from it are under the sea. It is conceivable that

[45] E. M. Kindle, "Observations on Ice-borne Sediments by the Canadian and Other Arctic Expeditions," *Amer. Jour. Sci.*, 5th ser., Vol. VII (1924), pp. 251–86, 264.

floe ice scouring a cave-pitted limestone coast could enclose and later transport large masses of rock, but it is probably much less competent for this work than would be an ice cap.

Powers'[46] theory should be especially noted because under it shorter transportation is required. According to him, a land existing in Caney time, on which were exposed the requisite Arbuckle-like rocks, was located in the present site of the Ouachitas, and was much later covered by northward overthrust sheets of Ouachita rocks which had been deposited farther south off the shores of Llanoria. Powers has stated some of the arguments against the theory. Boulders were supposedly derived from the land of Arbuckle-like rocks, but transportation of many miles (at least 10 miles) would have been required to bring them southward into Ouachita deposits.

If floe or berg ice rafts got such masses of rock into them, which seems improbable, and got into open seas, distance of travel is not an important matter, for they might have had clear sailing for many miles before breaking up.

Scratched and grooved "Caney" boulders have been found at a few localities. Compton cut is the best-known locality and the only one known at which scratched boulders are abundant. Woodworth[47] who has studied this locality has shown rather conclusively that the grooves are of tectonic origin resulting from one boulder grazing another during diastrophism subsequent to the deposition of the boulders in the shale. Scratches on boulders at other localities are probably of the same origin. As the "Caney" shale is an incompetent, abundantly drag-folded formation in the Ouachitas, one may expect to find scratched boulders in many localities where boulders are close together in their horizons.

Powers[48] and Miser[49] have considered the possibility of southward overthrusting during Caney time of Arbuckle-like rocks over Ouachita rocks. If the Arbuckle-like rocks were thrust southward in this manner, the gliding planes must have been in the Caney shale, because faults such as those postulated are not known in the older rocks. The blocks and boulders would be erosional remnants of such an overthrust mass. Southward overthrusting is not the direction of

[46] *Op. cit.*, p. 1044.
[47] *Op. cit.*
[48] *Op. cit.*, p. 1046.
[49] *Op. cit.*, p. 29.

movement which might be expected, if one believes that the later, post-Boggy, diastrophism which folded the Ouachitas was directed from the south. Facts suggesting that the direction of this mountain-making diastrophism was from the south or southeast are: (1) that the folding becomes less intense toward the north, (2) that the folds in the northern part of the mountains are overturned toward the north, and (3) that the northwest arc of the mountains forms a salient projecting toward the northwest. There is considerable doubt of the competency of thin shells of the earth's upper crust to transmit pressures for great distances. In McCurtain County on the south flank of the mountains the folds are overturned toward the south. As much of the south flank of the fold belt is concealed under the coastal plain, this southward overturning is the only recognized evidence indicating that the Ouachitas constitute a wedge-shaped sector which has been folded by compression from both the north and the south, rather than that the mountains were folded by pressure directed only from the southeast. With this view in mind, while reasoning from analogy, it is unsafe to conclude that Arbuckle-like rocks which were deposited to the north of the present Ouachitas could not have been thrust southward over Ouachita rocks during Caney time. However, the Caney boulders themselves are the only, and not very satisfying, evidence of this hypothetical southward overthrusting.

Dake[50] first suggested that "an enormous overthrust fault, the displacement measured in scores of miles—may have crowded the Ouachita facies from the south or southeast far to the northeast, over the Arbuckle facies. . . . ." This theory has been considered by Honess[51] in explaining the sources of the Ouachita asphalt deposits and by Powers[52] and van der Gracht[53] in explaining the origin of the "Caney" boulders. Any theory of long-distance overthrusting must necessarily involve low-angle faults. Miser[54] has added credence to the view that great, low-angle faults exist in the Oklahoma Ouach-

[50] C. L. Dake, "The Problem of the St. Peter Sandstone," *Mo. School of Mines and Metallurgy Bull.*, Vol. VI, No. 1 (1921), p. 55.

[51] "Geology of Atoka, Pushmataha, McCurtain, Bryan and Choctaw Counties (Oklahoma)," *Okla. Geol. Surv. Bull. 40-R* (1927), p. 26.

[52] *Op. cit.* [53] *Op. cit.* [54] *Op. cit.*, p. 19.

itas by his map of the areal relations in the Potato Hills which suggest a fault window. The belt of supposed overthrust rocks which bounds the possible window on the northwest side of Potato Hills is so narrow (less than 3 miles in width) as to cast some doubt on the probability that a section from the top of the Atoka through the Caney, Jackfork, and into the Stanley formation (a thickness presumably greater than 2 miles) is concealed under it. The writer has drawn a structure section (Fig. 3) illustrating the grounds for this doubt. Other interpretations of the areal relations in the Potato Hills are explained below.

The Boktukola fault cuts across the Boktukola syncline in northwestern McCurtain County. Miser[55] has attempted to measure horizontal displacement along the fault by measuring the displacement of presumably identical synclinal axes. It should be pointed out that two synclinal axes in their present positions on one side of the fault trace very nearly coincide with two others on the opposite side.

It may be unnecessary to postulate a hypothetical window[56] exposing older Paleozoic rocks in the Choctaw anticlinorium to explain an apparent lack of structural conformity within and outside of the supposed window. The older Paleozoic rocks are nearer a probably competent basement of pre-Cambrian rocks than are those of younger Paleozoic age, and folds are consequently more open with stratigraphic depth.

In the Arkansas Ouachitas the comparatively few faults are reported[57] to be thrusts of the broken-fold type, and there are probably many thrust faults in Oklahoma which are the result of the post-Boggy orogeny. In Oklahoma, however, there are many long faults which are subparallel to the grain of the country. Several extend the entire length of the mountains in Oklahoma from the coastal plain into western Arkansas. Since the time of Taff's[58] early work, these faults have been thought to be thrusts, although of high angle. Exposures of the surfaces of these faults have not been reported, and apparently have not been found.[59]

[55] *Ibid.*, p. 20.

[56] *Ibid.*, p. 22.

[57] *Ibid.*, p. 18.

[58] *Op. cit.*

[59] In a letter dated September 8, 1930, Mr. Miser stated that he knew of no locality in Oklahoma showing the exposure of such a fault surface.

The apparently unanimous belief that these long faults are thrusts is largely based upon regional and, to a certain extent, upon local structural relations. (1) As the diastrophism which produced the long faults was undoubtedly great, it has generally been considered to have been a part of the major, post-Boggy orogeny, and hence is supposed to have produced the crustal shortening of thrust faults. (2) The upthrow sides of the long faults are on the south, that is, on the sides from which the direction of pressure is thought to have come during the major orogeny, as should be the case if the faults are thrusts resulting from that diastrophism. (3) The long faults are subparallel to the grain of the region, as should be expected of thrusts produced by the same forces which produced the grain.

These general relations support the belief that the long faults are thrusts, as does the following more specific relation. The relations along most of the long faults are not well known, because valley-forming Stanley shale crops out along almost the entire lengths of their upthrow sides which are on the south. The Jackfork sandstone or other rocks at most localities on the downthrow sides dip steeply (about 40°–60°) southward toward the upthrow sides. These steep dips of the rocks on the downthrow sides toward the traces of the faults suggest that the faults are high-angle thrusts, although they do not preclude the possibility that the faults are normal.

The structural relations along the Choctaw fault, the longest of all, which crops out along the north-northwest border of the Oklahoma Ouachitas, are very different from, and are much better known than, the relations along the other long faults. Long and narrow outcrops of the Wapanucka limestone are cut off by this fault. This circumstance facilitates the study of the structural relations. The upthrow side of the fault is on the south, but on this side a steeper dip of about 60° away from the fault's trace usually prevails than the dip of about 30°–40°, also away from the trace, on the north or downthrow side. These relations not only indicate that the fault is of high angle but also suggest that it may be normal, because if the fault were a thrust resulting from the breaking of an overturned fold the strata on the upthrow side should dip less steeply than the beds on the downthrow side, which in many localities might be expected to be overturned. In the quarry at Stringtown, strata of the String-

town formation dip steeply westward on the east or upthrow side of the Choctaw fault, and on the west side Atoka strata dip steeply westward. These relations also suggest normal faulting.

If one assumes that the long faults occurred at the time of the folding, he must conclude that they are thrusts, but if he assumes that they occurred during relaxation subsequent to folding or during some later period of tension, he may well conclude that the faults are normal. In considering the possibility that the long faults may be normal, the uniformity with which downthrow sides are on the north may be explained by the theory that faulting was down from and around the old positive or competent area, Llanoria, which lay to the south. Normal faults of many miles length are known which are subparallel to a pre-existing grain. It should be noted that Honess found a number of normal faults in McCurtain County, and one of them is the only fault[60] whose dip could be measured. Proof that the long faults are normal would add credence to Robinson's[61] suggestion that there is a genetic relationship between them and the Balcones and Mexia fault zones of Texas. The Texas zones may have been faulted with downthrow sides away from Llanoria during the same epoch as were the long faults of the Oklahoma Ouachitas, and during post-Cretaceous time, after the cessation of the faulting in Oklahoma, additional displacements may have occurred along the older zones. In the case of the Mexia zone the downthrow was away from the still competent area which was the former positive element, Llanoria, but in the case of the Balcones zone the downthrow was dominantly away fron the Central Mineral region of Texas, thus producing the graben between the Mexia zone on the southeast and the Balcones zone on the northwest. Possibly pre-Cretaceous Balcones and Mexia normal fault zones connected with the long faults in the Ouachita Mountains.

While the writer is far from being entirely convinced that the long faults are normal, the bulk of the evidence favors the conclusion that they are of high angle. In the absence of more positive evidence, the

[60] Honess, "Geology of the Southern Ouachita Mountains of Oklahoma," *Okla. Geol. Surv. Bull. XXXII*, Part I (1923), p. 241.

[61] H. M. Robinson, "The Origin of the Structure of Northeast Texas Petroleum Area," *Econ. Geol.*, Vol. XVIII (1923), pp. 722–31, 726.

steep dips adjoining the traces of the faults indicate that the surfaces of the faults are likewise steeply inclined. Furthermore, these faults do not show the usual characteristics of low-angle, long-distance overthrusts. Windows, if present, are a criterion of low-angle thrusting, but they are much less common in regions which are known to have this type of faulting than are either outlying klippen or peninsula–re-entrant outcrop patterns of the overthrust masses, whether these patterns are of the dendritic type associated with a nearly plane fault surface or of the type associated with an undulatory surface as in the case of the Rome fault in Georgia. At no place along the several hundred miles of long fault outcrops do such phenomena occur. The lack of such evidence of low-angle faulting casts doubt upon the actuality of windows. If the surfaces of such hypothetical overthrust faults are thought to have been folded during or after thrusting, so that they dip steeply in most outcrops, surely in many places along anticlinal axes of the folded fault surfaces, these surfaces should be nearly horizontal, and should show peninsula–re-entrant outcrop patterns of the overthrust sheets.

BOULDER FACIES CONSIDERED AS NON-EXOTIC

Although older Paleozoic rocks in the "Caney" boulders are not like outcropping Ouachita rocks, but are more like Arbuckle rocks, boulders of younger Paleozoic age are composed of Ouachita rock. An explanation of the younger boulders does not require long-distance transportation, and the evidence afforded by some of them appears to preclude such transportation. The friable and soft shale boulders in the sandstone at Stapp could not have been transported far by any process which might have crushed or shattered them. Blocks of Caney shale in modern stream channels are either on or within a few hundred feet of an outcrop of the parent rock. A friable boulder of chert breccia found in southeastern Atoka County would probably have been shattered during any considerable transportation. The great masses, or boulders, of older Paleozoic rock in the shale in Johns Valley have probably not been transported any considerable distance. It is a reasonable assumption that all of the boulders have been derived from the same general source.

As no probable extraneous source of the boulders is known, the

Ouachita area itself should be considered as a possible source. All of the boulders may have been derived from local rocks, but the older Paleozoic, Arbuckle-like section does not crop out. The older Paleozoic Ouachita rocks in Oklahoma are confined in outcrop to McCurtain County and to the Potato Hills and Black Knob Ridge inliers. The Potato Hills are 30 miles northwest of the McCurtain County outcrop, and Black Knob Ridge is 40 miles west-southwest of the Potato Hills and 65 miles west-northwest of the McCurtain County area. Because of crustal shortening, these intervening distances are less than they were at the time of deposition of the exposed Carboniferous rocks. Because the only outcropping, older Paleozoic rocks in the Black Knob Ridge and Potato Hills inliers are similar to those in the McCurtain County area, it has been assumed in the past that rocks of the same facies underlie the intervening area.

Drill holes in this intervening area have not encountered rocks of greater age than Carboniferous, nor have pre-Carboniferous rocks been reached in wells to the southwest of it on the coastal plain in Oklahoma. Wells, however, drilled farther east, on the coastal plain in Oklahoma and adjacent Texas[62] to the southwest of the McCurtain County inlier of pre-Carboniferous rocks, have penetrated older Paleozoic and Devonian rocks of the outcropping Ouachita facies. Other wells in the vicinity of Denison, Texas, 45 miles south-southwest of the west end of the Ouachitas, have penetrated similar pre-Carboniferous strata. But, in the heart of the Oklahoma Ouachitas there is a diamond-shaped area, under which the nature of the pre-Carboniferous rocks is not known. The area is 100 miles long by 40 miles in greatest width, the long axis of the diamond extending in an east-northeast direction.

There is some probability that Arbuckle-like, older Paleozoic rocks were deposited in part of this area. A well drilled on the west side of the area penetrated a normal section of Arbuckle rocks from "Carboniferous Caney shale down to the Ordovician Simpson formation"[63] after drilling through a thin veneer of Cretaceous. This

[62] H. D. Miser and E. H. Sellards, "Pre-Cretaceous Rocks Found in Wells in Gulf Coastal Plain South of Ouachita Mountains," *Amer. Assoc. Petr. Geol.*, Vol. XV (1931) pp. 801–18.

[63] *Ibid.*, p. 808.

well (Hansen *et al.*, in the SE. ¼, NW. ¼, Sec. 17, T. 4 S., R. 11 E., Atoka County, Oklahoma) is located on the coastal plain, 10 miles south-southwest of Black Knob Ridge, about 2 miles east of the southwesterly projection of the strike line of the pre-Carboniferous rocks in the ridge. Thus, it appears to be located within the concealed Ouachitas near their west end.

A Trinity (Cretaceous) limestone conglomerate or breccia which crops out well within the critical area further suggests that Arbuckle-like, pre-Carboniferous rocks had been deposited nearby, and had outcrops in Cretaceous time which were the source of the limestone fragments. This conglomerate crops out (NE. cor., Sec. 5, T. 4 S., R. 14 E.) 4 miles west of Darwin in southeastern Atoka County, Oklahoma. This location is only a few miles south of several "Caney" shale boulder localities. Small (up to 3 cm.), angular to rounded, much-weathered, gray to buff limestone fragments are in a friable sandstone matrix. Caney boulders, whose horizons cropped out at the time of deposition of the conglomerate, were probably not an adequate source of the abundant limestone fragments. If the fragments had been derived from a monadnock in which Caney boulders cropped out, probably quartzitic Jackfork sandstone also would have been cropping out in the same erosional remnant, and fragments of this more resistant rock undoubtedly would have been more abundant than those from limestone boulders. There are Trinity limestones which are older than the conglomerate, but they are not present in the Cretaceous section[64] near this locality in which younger Trinity horizons overlap the truncated Paleozoics. The fragments in the limestone conglomerate were probably derived from a very near outcrop of Paleozoic limestone. It should be noted that this outcrop of the conglomerate is located only a short distance west of the projected axis of the syncline in the center of which Johns Valley is located.

If the rocks in the boulders are taken to represent a section, most of Hunton time is represented by a hiatus, and much of this recordless time also corresponds to hiatuses[65] in the Talihina chert. As the

[64] H. C. Vanderpool, "A Preliminary Study of the Trinity Group in Southwestern Arkansas, Southeastern Oklahoma, and Northern Texas," *Amer. Assoc. Petr. Geol. Bull.*, Vol. XII (1928), pp. 1069–94, Figs. 2 and 3.

[65] Ulrich, *op. cit.*, p. 30, Fig. 2.

four Hunton formations above the Chimneyhill in the Arbuckles are fossiliferous, it is probable that they are not represented in boulders or they would have been identified, and accordingly these formations were probably not present in the section of rocks from which the boulders were derived. Lower Simpson time also appears to be represented by hiatuses both in rocks in boulders and in outcropping Ouachita rocks. A correlation of these hiatuses suggests that the deposition area of the rocks which were the source of the older Paleozoic boulders had a physical history more comparable to the Ouachitas than to any other known area.

It is probable that Arbuckle-like rocks of older Paleozoic age underlie the Arkansas Valley in Oklahoma. It has been reported[66] that a well (Pattison and Phillips, McDuff farm, Cen. NW. $\frac{1}{4}$, SE. $\frac{1}{4}$, Sec. 32, T. 9 N., R. 16 E., Pittsburg County, Oklahoma), located about 20 miles north of the Ouachitas, encountered probable Viola limestone at a depth of 5,268 feet. Ulrich[67] has postulated a "former topographic boundary" along the northwest side of the Ouachitas between Ouachita rocks and Arbuckle-like rocks. As rocks of younger Paleozoic age like those which crop out along the north side of the Ouachitas were the source of many boulders, it is probable that older Paleozoics from, or near, the same region were the source of others. Rocks like the Middle Devonian cherty beds on Brushy Creek are known from nowhere else in Oklahoma except at this Ouachita locality.

BENGALIA, THE LAND AREA WITHIN THE OUACHITAS

A logical theory is required to explain how an older Paleozoic, Arbuckle-like section of rocks might have been deposited in the diamond-shaped area between McCurtain County and the Potato Hills and Black Knob Ridge.

There is evidence[68] that land existed along the north side of the Ouachita area in Arkansas during several periods of pre-Stanley Paleozoic time, and hiatuses in the section of this age which is ex-

[66] W. W. Clawson, Jr., "Oil and Gas Geology of Coal and Pittsburg Counties (Oklahoma)," *Okla. Geol. Surv. Bull. 40-JJ* (1928), p. 15.

[67] *Op. cit.*, p. 26.

[68] H. D. Miser and A. H. Purdue, *U.S. Geol. Surv. Geol. Atlas, Hot Springs Folio No. 215* (1923); "Geology of the De Queen and Caddo Cap Quadrangles, Arkansas," *U.S. Geol. Surv. Bull. 808* (1929), p. 44.

posed in the Potato Hills and Black Knob Ridge indicate that there were land areas along the north side of the Oklahoma Ouachitas. At times this land, for which the name "Bengalia" is here proposed, was composed of a few low islands, whereas during other periods it formed a continuous low barrier which may have connected with Llanoria, the land to the south. Early in the Paleozoic, pre-Cambrian rocks were exposed on Bengalia, and later, as a result of changes in the positions of land and sea areas, younger rocks formed the land. Bengalia probably tended to retain its position along the north side of the Ouachita area, because of the tendency of a basin of deposition to retain synclinal structure when warped by tangential pressure.[69] The central basin, however, was probably land at various times, as testified by hiatuses revealed by the rocks in the "Caney" boulders.

In pre-Stanley time streams from Llanoria and Bengalia transported colloidal silica to the sea, where it was precipitated near shore. For long periods of time the lands were so low that the seas were clear and allowed the deposition of relatively pure bedded cherts, although pre-Stanley shales and sandstones aggregate at least twice the thickness of the chert beds. As silica[70] is evidently precipitated almost immediately upon entering the sea, such chert beds were deposited very near the lands and probably do not underlie all of the central area between Llanoria and Bengalia. In this area, which evidently connected with the Arbuckle sea by way of a strait south of Black Knob Ridge, Arbuckle-like rocks were deposited. Older Paleozoic Bengalia lands were probably located to the east of the Potato Hills and Black Knob Ridge inliers. In Arkansas (novaculite) time Bengalia probably extended most of the length of the Ouachitas. The intervening area between Bengalia and Llanoria was so narrow in Arkansas during pre-Stanley time that the chert facies covered the basin.

Bengalia, with Llanoria, probably had its climax as a land area during post-Arkansas (novaculite) pre-Caney (Mississippian) time. Llanoria became high and supplied most of the clastic material for the Stanley-Jackfork succession, and Bengalia rose and formed a low

[69] Bailey Willis, "Mechanics of Appalachian Structure," *U.S. Geol. Surv., 13th Ann. Rept.*, Part II (1893), pp. 217–81, 262.

[70] W. H. Twenhofel, *Treatise on Sedimentation* (Baltimore, 1926), p. 383.

barrier on the north side of the filter-basin-like trough which received the sediments. Bengalia extended far to the south and east so that a thick Stanley-Jackfork succession was confined to the narrow trough at the same time that only a few hundred feet of lower Caney shale were deposited to the north and west of the barrier. Arbuckle-like rocks as well as strata of the chert facies cropped out in Bengalia. In the west end of the Ouachitas a land area of tilted Arbuckle-like rocks extended along the axis of the Johns Valley syncline, rising above the floor of the basin. The rocks in the Potato Hills area had been warped into a gentle dome which had collapsed with normal faulting before Stanley deposition. As the sea encroached upon this low area, the Stanley shales, sandstones, and chert conglomerate[71] were deposited between islands of older rocks. The areal distribution of rocks in the Potato Hills may also be explained by the fault-window theory,[72] which has already been discussed, or by the theory of the collapse of a dome with normal faulting subsequent to post-Boggy folding of the Ouachitas, but the evidence of land during Stanley time, as testified by the chert conglomerate, favors the theory of pre-Stanley warping and normal faulting. Typical conglomerates are relatively rare in the sediments deposited along the shores of Bengalia probably because the land was low and streams were short.

BOULDERS FROM BENGALIA

By later Caney (Johns Valley) time Llanoria had been eroded lower and Bengalia was a chain of islands with many straits which connected the Ouachita basin with the sea to the north and west. Caney shale was deposited in the north side of the basin and continued to be deposited to the north and west of Bengalia. Nearer Llanoria, in the basin in Arkansas and in the south side of the basin in Oklahoma, coarser clastics continued to be deposited throughout Caney and into Atoka time.

Early in this Johns Valley epoch of Caney time, the Ouachita basin in Oklahoma began to sink, probably accompanied by normal faulting along the south side of the Bengalia islands. The strata exposed in Bengalia had already been warped, faulted, and eroded in

[71] Powers, *op. cit.*, p. 1039. [72] Miser, *op. cit.*, p. 20.

several preceding periods so that rocks ranging in age from Ordovician to Mississippian (about 6,000 feet of section) were exposed.

Miser and Purdue[73] have noted that the Blaylock sandstone (Silurian) and older rocks in Arkansas are more crumpled than the overlying strata. Perhaps this deformation is the record of the first diastrophism which warped Bengalia, but later movements in Devonian, and especially in early Mississippian, time largely contributed to the exposure of this section of rocks. At the contact between the Arkansas novaculite (dip 73°) and the Atoka (dip 55°), 1 mile west of Bengal, the chert dips 18° more steeply than does the younger Atoka shale.

Earth tremors accompanying the downwarping of the basin, or other causes, may have produced subaerial and submarine landslips of fragments of Bengalia rocks into the Caney shale then being deposited. Subaerial processes had probably transported many of the smaller fragments into or near the sea, but submarine landslips probably were responsible for the wider distribution of these "Caney" boulders. Submarine landslips have been postulated as a process capable of transporting large masses of rock,[74] but, because such modern slips occur in the sea, distance of carriage and size of blocks transported have not been determined. Mud flowage[75] is the subaerial process probably most nearly analogous to submarine land slippage. Mudflows have been observed to carry great rock fragments down relatively gentle slopes. Subaerial mudflows, however, cease to move after becoming too viscous from the loss of water, and must dam up and absorb more water before they can flow again.

Submarine mudflows resulting from landslips, having abundant water, would probably flow continuously until a very low angle of rest (perhaps as low as 2° or 3°) is reached. Five houses which slipped into Lake Limon[76] at Zoug, Switzerland, on July 5, 1887, came to rest

[73] Miser and Purdue, "Geology of the De Queen and Caddo Cap Quadrangles, Arkansas," *op. cit.*, p. 44.

[74] E. B. Bailey, L. W. Collet, and R. M. Field, "Paleozoic Submarine Landslips near Quebec City," *Jour. Geol.*, Vol. XXXVI (1928), pp. 577–614.

[75] Sir W. M. Conway, "Exploration in the Mustagh Mountains," *Geog. Jour.*, Vol. II (1893), pp. 280–303, 291.

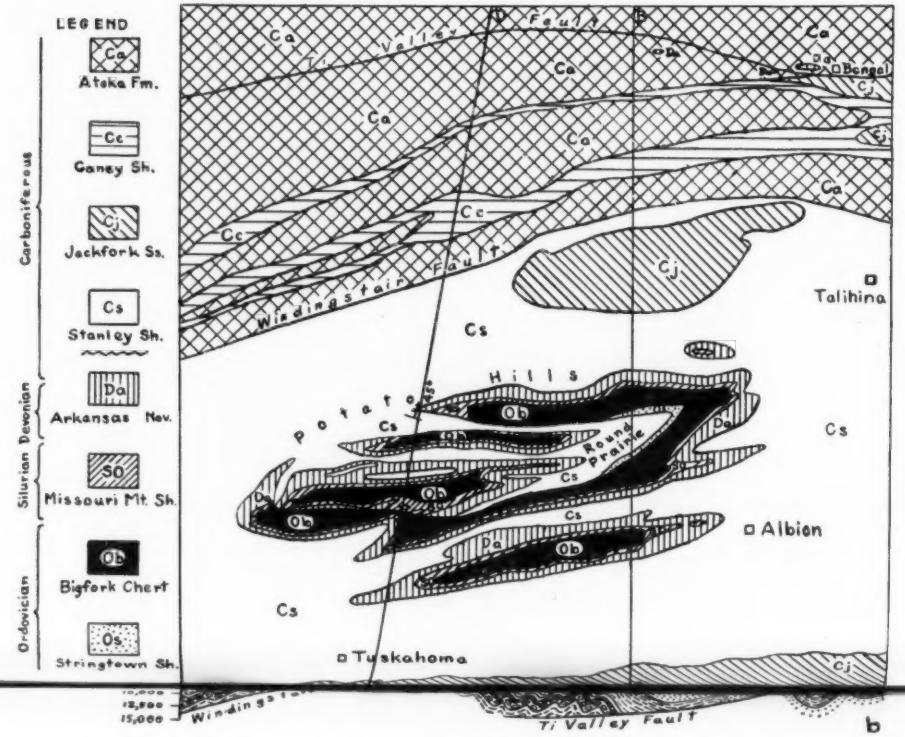
E. Blackwelder, "Mudflow as a Geologic Agent in Semi-arid Mountains," *Geol. Soc Amer. Bull.*, Vol. XXXIX (1928), pp. 465–84.

[76] L. W. Collet, *Les Lacs* (1925), p. 194.

at a depth of 45 meters at a distance of 1,020 meters from, and about $2\frac{1}{2}°$ below, the point of departure. It is conceivable that in a flow of considerable magnitude even huge masses of rock might be transported several miles.

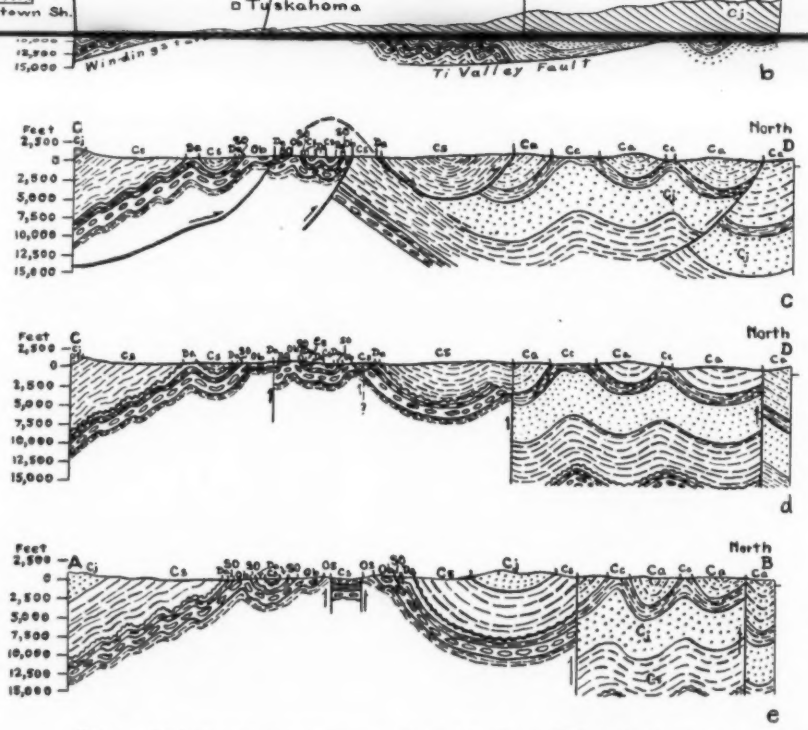
An outlying Bengalia island located in the Johns Valley area may well have been the source of the immense Caney boulders in that locality. The blocks are of such great dimensions that it is improbable that any ordinary sedimentary process has transported them far, and there is no evidence that they were tectonically transported. The Arbuckle-like rocks projected up through the partially overlapping Stanley and Jackfork until buried by upper Caney and Atoka strata. On first thought it would seem that the highest point on the island must have had an elevation about 12,000 feet above the floor of the basin in which these two thick formations were deposited. But it is more probable that the measured thicknesses of these formations are only apparent and not true, and that nowhere was the Stanley-Jackfork succession deposited in a thickness greater than a fourth to a half of the measured thickness of these strata. The Ouachita trough probably had an uneven floor with minor depressions and elevations upon it. The Stanley-Jackfork sediments were laid down with depositional dips into the depressions and away from the elevations. When the area was subsequently folded, the sediments in the minor basins of deposition became synclinoria and those on the elevations became anticlinoria.

Sediments preserved in one of these gently folded synclines afford the best section for measuring apparent thickness. The Bengalia island in the Johns Valley locality need not have had an elevation of more than 3,000–4,000 feet above the lowest point on the floor of the major basin. The island was probably a detail rising above the floor of one of the secondary basins and caused depositional dip away from it only over a minor area. For this reason, the buried hill, at the time of diastrophism, only determined the location of the center of the syncline instead of the position of an anticline. Had the buried hill been of considerable length, and elongated approximately at right angles to the direction of compression, it probably would have determined the location of an anticline, but if the hill was long and elongated approximately parallel to the direction of pressure, it probably would have buttressed the strata and determined the posi-



a) Map showing areal relations (after

c) Structure section along line, *C–D*, based on the fault-window theory.



a) Map showing areal relations (after [illegible]).

c) Structure section along line, *C–D*, based on the fault-window theory, showing a minimum thickness of strata which must be covered by the narrow belt of supposed overthrust rocks on the northwest side of the Potato Hills. This thickness is so great that Jackfork sandstone should crop out within the window. If the strata have undulatory structure and if the Atoka formation is partially repeated under this part of the supposed overthrust sheet, the covering would by no means be wide enough to conceal the section from Stanley to Atoka.

d) Structure section along line, *C–D*, based on the theory of Stanley overlap subsequent to post-Arkansas (novaculite) doming and faulting.

e) Structure section along line, *A–B*, based on the theory of Stanley overlap subsequent to post-Arkansas (novaculite) doming and faulting.

FIG. 3.—Geologic map and structure sections of the Potato Hills and adjacent areas in southeastern Oklahoma

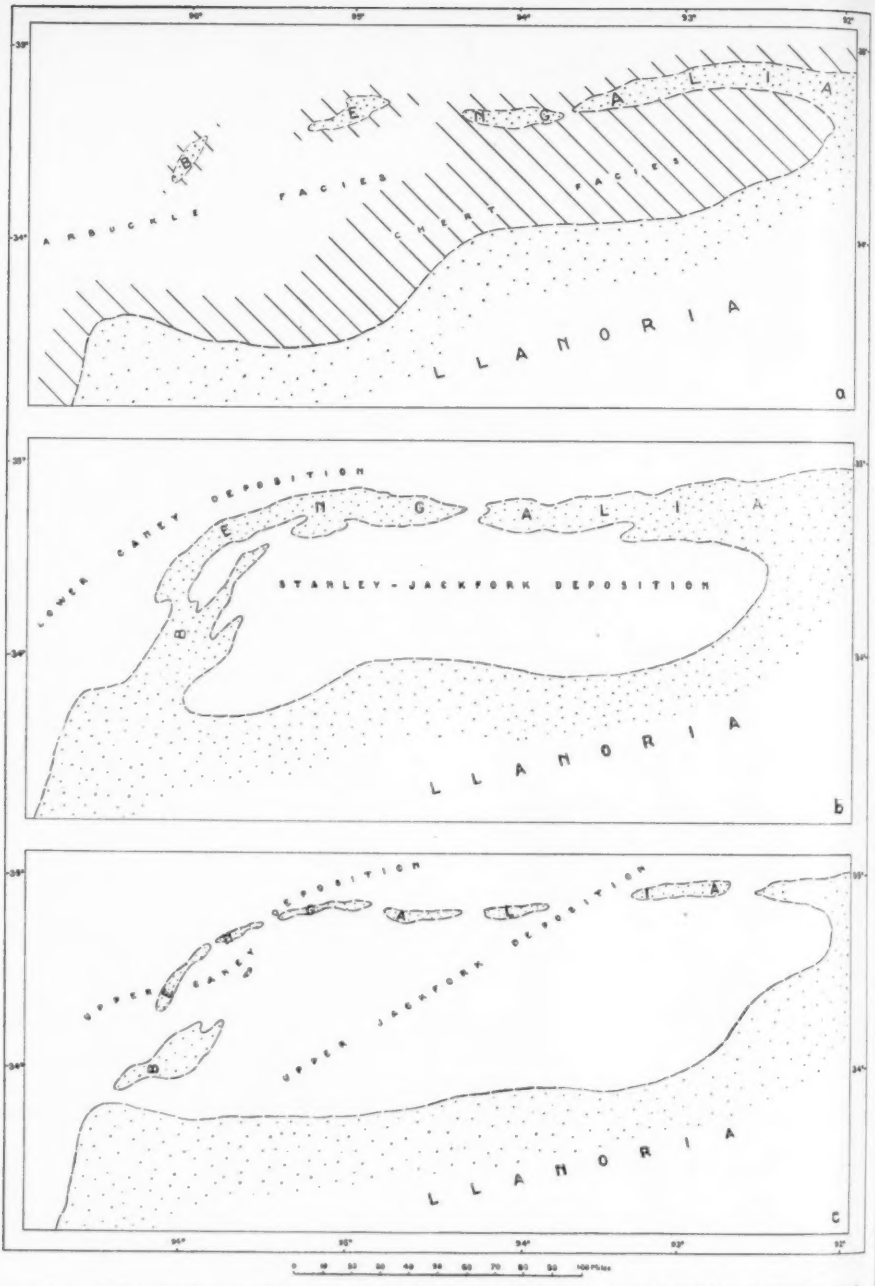


FIG. 4.—Paleographic maps of the Ouachita region during Paleozoic time. *a*, Ordovician; *b*, Stanley-Jackfork, or early Caney (Mississippian); *c*, Johns Valley epoch of later Caney (Mississippian and Pennsylvanian).

tion of a syncline. In this position the buried hill would have caused upwarping of the strata at the end nearer the direction from which pressure was transmitted, by resisting pressure, and in transmitting pressure would have caused the strata, at the opposite end to buckle into a probably arcuate, anticlinal fold. The writer performed pressure-box experiments with a number of buried-hill models, some of which are illustrated in Figure 6. Conclusions regarding the effects of variously oriented, buried hills in locating anticlines and synclines in folded strata are taken from these experiments.

Boulders of more than 30 feet in greatest dimension are not known outside of Johns Valley, perhaps because most of the outcrops of

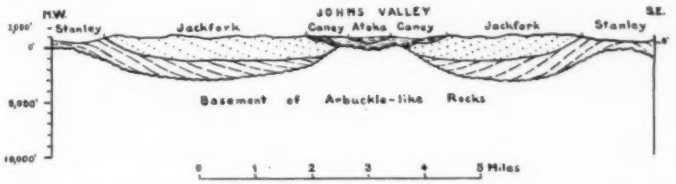


FIG. 5.—Structure section across Johns Valley from northwest to southeast (from S.E. cor., Sec. 14, T.1 N., R. 15 E., to S.E. cor. Sec. 17, T.1 S., R. 17 E.).

boulder horizons are farther seaward from the sources of rock fragments, or possibly represent boulders which flowed down gentler slopes, or were farther from earthquake centers, and flowed as the result of less intense tremors.

The inliers of pre-Stanley rocks (Black Knob Ridge, Wesley, Brushy Creek, 7 miles south of Hartshorne, and the two west of Bengal) within the Atoka belt on the north-northwest side of the mountains are approximately on the axis of Bengalia as it existed from Johns Valley Caney time to early Atoka time. The chert rocks, being better ridge-makers than the Arbuckle-like rocks, maintained higher elevation, and composed the higher areas of Bengalia which were last to be overlapped and which form the only present-day inliers of the former land. The main belt of boulder-horizon outcrops is a few miles south of this probable axis. Of the six known inliers of "Caney" strata along the axis—at Wesley, Brushy Creek, 2 miles northeast of Ti, two localities southeast of Recyl, and 8 miles east of Bengal—the three last mentioned contain boulder horizons. As

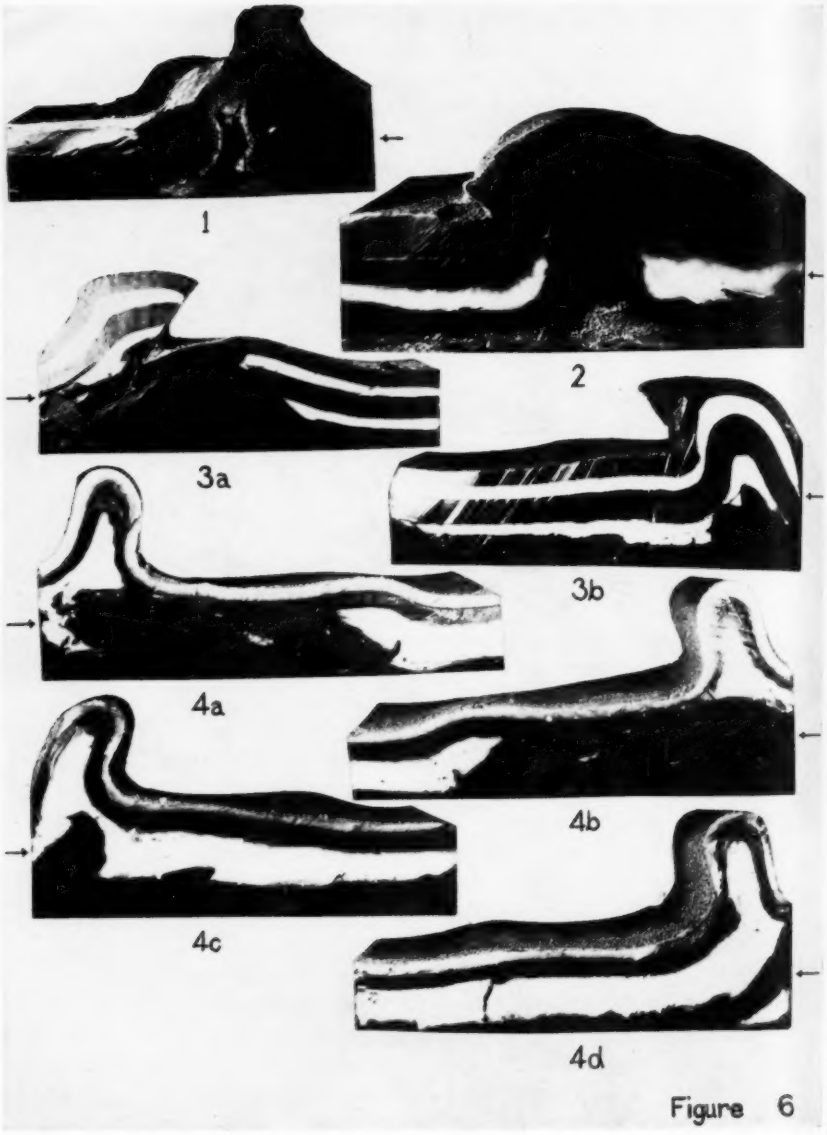


FIG. 6.—(legend on opposite page)

the horizons are not continuous as boulder beds, and as the areas of outcrop are small, three boulder-horizon outcrops in the six inliers are as many as might be expected.

Boulders were apparently not deposited in the Caney shale on the north side of Bengalia probably because the sea floor sloped gently northward from the land in that direction, and was less disturbed by crustal movements.

Boulders continued to be deposited intermittently from late Mississippian into early Pennsylvanian time when most of the remaining Bengalia islands were overlapped by Atoka strata. Perhaps Black Knob Ridge, which may have been rejuvenated by an early Pennsylvanian uplift,[77] was the last Bengalia island to stand above the transgressing sea; for there are conglomerates of Talihina chert and Stringtown shale fragments in the Atoka, Hartshorne, McAlester, Savanna, and Boggy formations a few miles to the north and west of this inlier.

Some time during the interval between Boggy and early Cretaceous time[78] the strata of the Ouachita area were folded and faulted and subsequently truncated. Post-Cretaceous erosion has sculptured the region into the Ouachita Mountains of today.

[77] Powers, *op. cit.*, p. 1048.

[78] Miser, *op. cit.*, p. 27.

FIG. 6.—Buried-hill models compressed in a pressure box. Models were 15 inches long by 8 inches wide, and were shortened to 10 inches. A mixture of half-paraffin and half-vaseline was used in the layers.

1. Longitudinal section showing a hill of sand at right angles to the direction of pressure which determined the position of the anticline.
2. Longitudinal section showing a hill of paraffin-vaseline at right angles to the direction of pressure which determined the position of the anticline.
3. *a–b*, Longitudinal section and the side view of a model having a paraffin-vaseline buried hill 7 inches long with a crest 1 inch high. Faulting occurred at the end of the hill toward the direction of pressure, and the hill was shortened and thickened, but pressure was transmitted, and a slight anticlinal flexure was produced at the opposite end of the hill.
4. *a–b*, Views of a longitudinal section. *c–d*, Side views. Dimensions of the hill are the same as in 3, but the bottom layer and the hill were modeled from sand saturated with the wax. An anticline developed at the end of the hill nearer the direction of pressure, and the hill transmitted pressure which caused an arcuate anticlinal flexure to develop at the opposite end.

TALUS BEHAVIOR ABOVE TIMBER IN THE ROCKY MOUNTAINS[1]

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ABSTRACT

This article discusses some features of talus behavior above timber line in alpine climates of moderate precipitation. The angles of rest of talus material generally range up to 36½° and very rarely exceed that figure. The steepness of talus slopes is dependent largely upon the coarseness of material; an empirical table gives the angles of rest characteristic of talus fragments of different diameters. The patterns of talus and vegetation on slopes are shown to be related to several factors, of which the degree of slope is most important. Ridges of loose rock that form at the feet of talus cones, but are separated from them by troughs, are also described and explained as accumulations at the downslope edge of snowbanks; anomalous angles frequently observed on talus piles are related in origin to these ridges. The rôle of water in talus accumulation is briefly discussed.

INTRODUCTION

To anyone engaged in geological studies that carry him above timber line in temperate climates of moderate precipitation, the accumulation and behavior of rock talus must constantly present problems of great interest. The writer has recently spent eight field seasons in different localities in the Western United States, including especially two summers in the mountains of central Idaho and four in the Mosquito and Sawatch ranges and nearby parts of Colorado. In all of these places much of the work was carried on above timber line. Though the studies were not primarily physiographic, the opportunities that they afforded for observing the processes of weathering were so intriguing that, incidental to areal and mine studies, considerable information on talus formation and behavior was obtained.

As data bearing on this process increased, the writer sought to compare his observations and those of other students in the field. In this country both Blackwelder and Bryan[2] have lately made valuable contributions to the study of weathering in desert regions,

[1] Published by permission of the director of the United States Geological Survey.

[2] Kirk Bryan, "Recent Work on the Phenomena of Arid Regions," *Zeitschr. Geomorphologie*, Band V (1930), pp. 225–27.

examples of which are cited. Others, notably Walther,[3] have written on the same subject in German. To his surprise, however, the writer found in print very little bearing on the general nature of talus behavior at alpine altitudes in regions of moderate precipitation. The only noteworthy exceptions are earlier articles on rock glaciers at high altitudes—articles which do not bear immediately on the problems here considered—and a few short contributions by European geologists.[4] He therefore offers the following observations in the hope that they will serve alike as a stimulus and a basis for further, more complete studies of the smaller landforms.

Though attention was first directed to such observations in the course of field studies in Idaho, the quantitative data presented below, as well as most of the generalizations, are based upon work carried on in Colorado, and most of the localities referred to may be found on the Leadville, Colorado, topographic map published by the United States Geological Survey.

THE ANGLE OF REST

What is the "angle of rest"?—The term "angle of rest" has never been adequately defined. It is loosely assumed by geologists to be the maximum angle of slope at which loose material will resist the force of gravity and remain stationary. It is generally admitted, however, that no slope is absolutely stable, for a very slow movement, measurable in decades if not in years, may actually be taking place. The slope assumed by loose fragments at any moment is a function of size and form of the fragments, the friction between the particles, the inclined component of the vertical force developed by gravity due to the slope of the substratum of the accumulating talus, and the viscosity of the surrounding medium (e.g., air or water). Dynamic geologists should seek an adequate definition of these factors, as well as an analysis of the meaning of "angle of rest."

It has been recognized that stability is affected by internal friction, for talus and rock waste will move in wet weather though re-

[3] Johannes Walther, *Das Gesetz der Wüstenbildung in Gegenwart und Vorzeit: Quelle und Meyer* (Leipzig, 1924).

[4] Helmut Kanter, "Junge Abtragungsercheinungen in den Teriärgebieten des östlichen Kalabrien und eines Teiles der südlichen Basilikata," *Zeitschr. Geomorphologie*, Band IV (1929), pp. 161–79.

maining immobile in dry weather. This statement is confirmed by the innumerable records of landslides that have taken place when the sliding material was wetted. In the field a rough attempt was made to evaluate this factor by artificially starting the "running" of a given talus pile and noting the length of time required for it to come to apparent rest, but the results were inconclusive.

A rough measure of the "permanence" or stability of a talus slope is also afforded by the growth of grasses or other small shrubs on it. The unsatisfactoriness of this criterion comes largely from the fact that the time required for grass to advance up a slope results in a lag: on account of the need for soil accumulation the slope may be essentially "at rest" several years before it is even scantily occupied by the grass and may well stand for a decade before moderately carpeted. In the writer's experience, however, vegetation above timber line cannot hold its own, but is covered by talus advance, when the slope is more than $35\frac{1}{2}°$. Obviously the rates of growth of vegetation and of talus supply are important qualifying factors.

Maximum angle.—By the "angle of rest," as applied to loose material on a slope, is meant the maximum downslope angle of the surface of the talus, as measured from the horizontal. Repeated observations by the writer have convinced him that the maximum angle of rest on bare slopes, no matter what the size of fragments, is 42°. A very few observations, taken in places where the blocks were of greatest size and angularity, gave the value mentioned, but these were so rare and the depth of talus-cover in such places so shallow as to necessitate throwing out this figure as a basis for most generalizations.

By far the greater number of maximum-angle observations was $36\frac{1}{2}°$. This figure, like all other slope readings here reported, was obtained by sighting along the strike of the slope and setting a Brunton clinometer while holding the edge of the instrument parallel with the slope.

Influence of particle size.—Repeated observations make possible a generalized table (Table I), which gives fragment sizes for angles of rest on slopes above timber in regions of active talus formation. These diameters given are obviously generalizations, since on most slopes the sizes vary widely.

Table I, when combined with other considerations, makes clear why the profile of a typical talus pile is steeper higher up, where most of the larger blocks remain, than farther down, where most of the smaller fragments are found. The angle of rest being a function of the size of the particles, the upper, less extensively weathered part of the slope is made up of coarser particles and stands at a steeper angle. By ordinary weathering the size of the fragments is progressively reduced downslope and hence the angle of rest becomes gentler downhill. At lower altitudes and on strictly colluvial slopes, where particle size becomes very small, the progressive difference in size downslope has been attributed to more rapid transport downhill of the finer particles,[5] but the large size of the individual fragments

TABLE I

Average Diameter in Inches	Average Angle of Slope in Degrees
12	35
6	35
4	32–34
2	31–32
½–1	26–31

would seem to preclude this explanation on alpine slopes such as those here discussed.

In opposition to the foregoing explanation, it may be argued that the dominant size of the fragments on a talus heap is determined by the size of the particles when they first break off. Ignoring exceptional large particles that are seen to have reached the bottom of the slope, a moment's consideration shows the fallacy of this idea, for, other things being equal, the greater the size of the fragment, the greater its momentum; this is because the mass varies as the cube of the diameter, the surface area (and consequently the friction) only as the square of the diameter. The larger fragments should therefore move farther down the slope and the lower parts of talus heaps should be coarser, not finer. Since the reverse is the case, however, it is at least highly improbable that the observed sizes of talus fragments represent the initial sizes when they fell from the cliffs above; the initial sizes have been modified by a later process

[5] B. K. Emerson, *Agricultural Geology* (New York: John Wiley & Sons, 1928), p. 211.

and the most likely process is comminution by weathering. Much the same conclusion is favored (though not absolutely proved) by the relatively uniform size of the fragments in upper and lower part of rock slides such as those reported by Howe from the San Juan Mountains and the famous landslide at Frank, Alberta.[6]

Relation of small to large particles.—On many slopes the larger blocks were observed to support on their upslope sides smaller particles for which the slope was abnormal. This gradually flattens the slope to angles lower than normal for the underlying particles. Thus, in an amphitheater at the head of the East Fork of the Arkansas north of Leadville, Colorado, blocks 4 feet in diameter composed a 36° slope, but smaller particles averaging 1 inch in diameter were gradually covering the larger blocks and producing a gentler mean angle for the slope as a whole.

THE RÔLE OF VEGETATION

The process and resulting patterns—On a grassy slope the downhill movement of talus particles is greatly impeded by vegetation. All stages may be seen, from slopes where vegetation is advancing against the talus to those on which it is the talus that is gaining the upper hand. The writer's attention was first directed to this struggle by Mr. B. S. Butler, of the United States Geological Survey. Careful observation of the patterns of talus and grass and small shrubs has afforded a means of interpreting this conflict.

When the advance of talus from above is opposed by the growth of vegetation, there is at first a tendency for the talus to pile against the individual plant tufts on the uphill side. If the tufts are widely spaced and talus development moderately rapid, the vegetation cannot long resist complete burial. But if the tufts are closely spaced, the pattern developed is similar to that of dunes elongated at right angles to the wind: the talus spreads sideways from each tuft and eventually forms ridges which follow the contours. These ridges themselves then serve as obstructions behind which more talus accumulates.

[6] See, e.g., Ernest Howe, "Landslides in the San Juan Mountains, Colorado, *U.S. Geol. Surv. Prof. Paper 67* (1909), Pls. X–XV; R. G. McConnell and R. W. Brock, "Report on the Great Landslide at Frank, Alta., 1903," *Can. Geol. Surv. Ann. Rept.* (1903), Part VIII, p. 9, and Pls. III, VIII, IX, X.

As the process is continued the pile that gathers upslope above a single tuft of vegetation or talus ridge grows high enough to be able to migrate across the top of the barrier. The talus now moves downhill in slender streams, and as this goes on similarly in other places the rock waste assumes what may be designated a "downslope" pattern, long lines of débris extending across the contours. Such a downslope pattern is indeed the more common one and is developed in other ways as well, which, however, cannot be discussed here.

Illustrations.—The features described are well exhibited on the southeast slope of East Ball Mountain 5 miles southeast of Leadville, on the north slope of Horseshoe Gulch near Leavick or Horseshoe (about 10 miles southeast of Leadville), and especially on the northwest slope of American Flag Mountain in upper Taylor Park, Gunnison County—all in Colorado.

At the Leavick locality the slope studied consists of alternating beds of shale and sandstone of Pennsylvanian age. These strike N. 40° W. and dip 45° NE., whereas the slope on which they outcrop strikes N. 80° W. and inclines 30°–32° SW. Massive beds of Mississippian limestone, outcropping beneath the Pennsylvanian, support the slope along its westward extent, so that its inclination there is reduced to 28°–30°. On the westerly part the lines of accumulated talus assume the "contour" pattern, but farther east, where the slope inclination increases to 32°, the "downslope" pattern predominates, as may be seen in Figures 1 and 2, and in such places the talus even forms continuous streams that threaten to wipe out all of the vegetation.

At the East Ball Mountain locality the change in pattern from "contour" to "downslope" is noted where the slope increases from 26° to 33°. In this instance there was active undermining of the grass on the downhill side.

But the most striking instance is at American Flag Mountain, a domelike, rocky mass composed of slabby Mississippian limestone in beds parallel to that slope on which the phenomenon was observed. Here the critical change in pattern shown in Figures 3 and 4 takes place where the slope breaks from 31° to 33°. Incursions of the "downslope" pattern leave long tongues of vegetation extending up-

slope and still uninundated. Finally, where the slope grows yet steeper, even this scanty cover disappears and the entire mountain side is covered by fragments of rock.

FIG. 1.—General view of north slope of Horseshoe Gulch near Leavick. The "downslope" pattern of talus here predominates.

FIG. 2.—Detail of "downslope" pattern of talus on north slope of Horseshoe Gulch. View obliquely, but almost directly up the slope

Causes of talus advance.—It should be stated that talus incursions of this sort are not everywhere from the head of the slope and are not all derived through exfoliation on the barren crest above timber line. Slopes with irregularities in gradient may show talus accumu-

lation of this sort at any place where the inclination steepens. The local killing-off of sparse vegetation may facilitate the development of waste by laying bare the rock surface. Or the gradual exfoliation

FIG. 3.—Profile view of northwest slope, American Flag Mountain; a slope of 31° rises from left to right. Vegetation forms patches or discontinuous lines parallel to the contours—the "contour" pattern.

FIG. 4.—Profile view of slope on American Flag Mountain. Slope rises at 33° from left to right. Vegetation forms bands extending up the slope—the "downslope" pattern.

of an erratic boulder may shed rock waste upon a slope on which vegetation and talus accumulation had previously been in equilibrium, throwing the balance in favor of talus movement. Illustra-

tions of each of these processes have been observed by the writer, which show that a certain pattern of accumulation does not necessarily persist in a given locality, but may be changed in response to factors, other than the slope of the hillside, that are subject to change within short time intervals.

NIVATION RIDGES

Descriptions.—A common, though by no means ubiquitous, feature where talus is accumulating above the zone of timber and heavy vegetation is a ridge of rock fragments, separated by a slight depression from talus at the foot of a cliff from which the débris has fallen. The ridge may be as much as 30 feet high and its inward and outward slopes have gradients corresponding to the angle of rest for the size of the component rock fragments. For this ridge the writer proposes the term "nivation ridge."

Depressions such as that described above are generally filled with snow in the early summer. In places several such ridges may be seen, subparallel and successively lower valleyward (Figs. 5 and 6).

Origin.—Such ridges have the following origin. After the winter's snow has disappeared from most of the landscape, large patches of it still lie between the cliff and the valley below and are often protected by the steep cliff above from the sun's rays and from evaporating winds until well into the summer. Talus blocks, breaking from the rock ledges, roll down over the snow and continue bounding toward its lower edge—a process that can be observed with frequency (see Fig. 6). If they come to rest on the snow, they gradually melt their way downhill as well as vertically downward into the snow, the downhill component being due to the readier melting of the thin wedge of snow that stands between the block and the atmosphere on its downhill side. Such talus fragments in time build up the ridges described. The trough between the ridge and the cliff represents the position of maximum depth of the snow.

That this explanation is correct is shown (1) by the freshness of the material in the ridge; (2) by the angle of rest, which characterizes slopes on both sides of the ridge; (3) by the average size of this material, which shows no more abrasion than does the talus against the hill slope; and (4) by actual observation of the process.

This explanation was briefly presented, but not further developed,

by Crawford.[7] It was also offered in brief but correct form by Howe.[8] The rôle of snow as an erosion factor was recognized by

FIG. 5.—View along a slope, southwest of Napoleon Pass, lower Taylor Park, Colorado. Cliffs of rock to left, flanked by talus, with nivation ridge in right foreground. Note the trough shown in profile.

FIG. 6.—Double nivation ridge on northwest side of Dyer Peak, Evans Gulch, near Leadville, Colo. Note the boulders on the snowbanks.

Matthes,[9] in his excellent studies in the Bighorn Mountains, but talus accumulation was not clearly related by that writer to snow

[7] R. D. Crawford, "Geology and Ore Deposits of the Monarch and Tomichi Districts, Colorado," *Colo. Geol. Surv. Bull. 4* (1913), p. 34.

[8] *Op. cit.*, p. 36.

[9] F. E. Matthes, "Glacial Sculpture of the Bighorn Mountains, Wyoming," *U.S. Geol. Surv. 21st Ann. Rept.*, Part II (1900), pp. 179–85.

banks. Bowman[10] recognized that talus is shed upon the snow from the cliffs above, but assigned most of the waste material associated with the snowbanks to the erosive action of the snow.

When overgrown with grass such ridges are scarcely recognizable. Their irregular, hummocky surfaces then strongly suggest landslide topography.

Superimposed talus fans.—A study of these ridges has afforded also an explanation for anomalous talus slopes. Such a slope is seen

FIG. 7.—Profile view of talus, showing anomalies in slope of surface on south wall of Bartlett amphitheater, northeast of Leadville, Colo. The talus slope is gentle at the left and steepens noticeably to the right of the middle.

on the south side of Bartlett amphitheater near the Climax Mine, Fremont Pass, 11 miles northeast of Leadville, Colorado (Fig. 7). Beneath cliffs of pre-Cambrian rocks the talus, in blocks ranging up to 6 inches in diameter, forms a slope of 25°. Farther down the slope flattens to 16° and within about 200 feet steepens again abruptly to 26°. Below it flattens gradually to the valley floor after the manner of typical talus piles. The changes downslope from 25° to 16° to 26° successively, however, are abnormal and are not accompanied by any recognizable variations in coarseness of material or by any source of added talus that, fed out upon the slope, might change the "head" of the pile and so alter the angle.

[10] Isaiah Bowman, *The Andes of Southern Peru* (American Geographical Society, 1916), p. 287.

A study of the north wall of the same amphitheater furnishes an explanation for these anomalies (see Fig. 8). Here Bartlett Peak drops off on its southeast slope over steep cliffs, but on its southwest declines by a gentler slope toward the valley. At the foot of the cliffs typical talus cones are built up, fronted valleyward by nivation ridges. On the southwest side the talus cones have grown until the talus has filled the trough between cone and nivation ridge—a process probably aided in this place by the early melting in the afternoon

FIG. 8.—Superimposed talus cone on north wall of Bartlett amphitheater. The cones in middle of photograph are simple and fronted by nivation ridges; to left of center is a large superimposed cone produced by the advance of talus over the top of the nivation ridges.

sun of the snow that would otherwise help in the continued building of the nivation ridge. The filling of the trough gives the break in the slope that was observed on the opposite wall of the cirque, and now some talus moves from the feeding cliffs directly out upon the steeper valleyward slope that represents the outer surface of the nivation ridge. Thus there is formed what may correctly be described as a superimposed talus cone.

TALUS GULLIES AND RILLS

That running water also plays an important rôle in distributing a part of what is commonly called talus must not be ignored. A usual feature of talus cones, even in such moist climates as prevail in the southern Rocky Mountains of Colorado, where cloudbursts are

rare, is a small, talus-floored trough, the walls of which rise slightly above the sloping surface of the bordering talus, like miniature lev-

FIG. 9.—Talus streams confluent toward clefts in rock cliffs, below which compound cones of larger dimensions project into the valley. East wall of Empire amphitheater 8 miles southeast of Leadville.

FIG. 10.—Torrential gully on one of the cones shown in Fig. 9 and small alluvial fan at its lower end.

ees. The writer has observed troughs of this kind as much as 8 feet in depth and 1,000 feet in length. At the lower ends of such troughs the débris is spread out as alluvial fans of delta-like shape. These features clearly represent the work of mountain torrents produced,

not by very sudden and heavy precipitation in small areas, such as is common in the Basin and Range region, but by the concentration of runoff through isolated and narrow clefts in otherwise unbroken rock walls. In times of heavy precipitation the torrents, rushing out through these rock clefts or miniature canyons, cut deeply into their talus cones and erode the trenches described above. It is at the lower ends of the trench that the water deploys as a sheet flood, distributing its load in the form of a typical alluvial fan of small dimensions, like that illustrated in Figure 10. A recognition of the importance of clefts in directing the torrents was arrived at independently by the writer in agreement with the conclusion recently published by Matthes.[11] As illustrated in Figure 10, this furnishes a means for rapid burial of vegetation and encroachment of talus upon a hitherto stable slope—another phase in the constant struggle between talus and vegetation.

[11] F. E. Matthes, "Geologic History of the Yosemite Valley," *U.S. Geol. Surv. Prof Paper 160* (1930), pp. 108–9.

THE ORIGIN OF MOUNTAIN LAKE, VIRGINIA

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ABSTRACT

Mountain Lake occupies an unusual position on the axis of a breached anticline, where two homoclinal ridges overlook it. The valley of a stream flowing through one of these ridges was apparently dammed by blocks creeping downward from the ridge, thus impounding the lake.

Lakes within the glaciated region of North America are such common features that their occurrence excites little notice unless they are of great size or of peculiar shape. Beyond the moraines, however, inland lakes other than those occurring in sinks or on the floodplains of mature streams are unusual, and, even when small, their origin is a source of speculation to layman as well as to geologist. When in addition to being extra-morainic the lake occupies a basin in a high valley surrounded by mountain ridges with whose development it is intimately related, the occurrence is unusual and invites explanation.

Such is Mountain Lake situated in Giles County 7 miles northeast of Pembroke in the Ridge and Valley section of western Virginia and shown on the Dublin, Virginia, topographic sheet. It was first brought to the writer's attention in 1923 by Professor George D. Hubbard, of Oberlin College, but in spite of its interest apparently no discussion of its origin has been published. The lake (Fig. 1) is $\frac{3}{4}$ mile in length and $\frac{1}{4}$ mile wide, having a surface elevation of 3,873 feet. As these dimensions are essentially the same as those given by a party of surveyors which is said to have discovered the lake in 1751, an artificial origin need not be considered. The location is one of rare beauty and is now the site of a small summer resort, which fortunately has not destroyed the natural charm.

Figure 2 shows that the lake occupies the axis of a breached anticline from the trend of which its longer direction diverges slightly. East of the lake the anticline is maintained unbreached by Silurian sandstones, of which the extremely resistant Clinch[1] is most im-

[1] G. D. Hubbard and Carey Croneis, "Notes on the Geology of Giles County, Virginia," *Denison Univ. Bull. Jour. Sci. Labs.*, Vol. XX (1924), pp. 307–77. All stratigraphic information is from this paper.

portant and forms the anticlinal Salt Pond Mountain, here about 4,300 feet in elevation. At the lake the resistant arch of sandstone has been breached, and Salt Pond Mountain divides, becoming Doe Mountain and Bald Knob, homoclinal ridges whose steeper scarp slopes oppose each other. The anticlinal valley thus formed is opened upon the Sevier shales, a formation here believed to be well over 1,000 feet in thickness and varying from calcareous shale at the bottom to sandy shale at the top. Due to the northeastward plunge

FIG. 1.—Northward across Mountain Lake from Salt Pond Mountain, the outlet occupies the low sag shown, but cannot be seen because of the projecting point.

of the anticline the two homoclinal ridges diverge toward the southwest and the intervening shale valley opens in that direction, where it is occupied by Doe Creek. The latter is a subsequent stream heading in a steep gully at a divide about 60 feet above the surface of the lake and about 600 feet from it. The lake occupies the anticlinal valley northeast of this divide and has no permanent tributaries except small spring rills. It does not parallel the anticlinal axis, but extends more to the north, appearing to fall in line with the outlet stream, a tributary of Little Stony Creek. Since the lake clearly lies within the anticlinal arch of Clinch, the outlet flows through and down a notch cut in this formation just north of the lake. The stream is resequent and is structurally superposed upon the Clinch. It has

a much lower gradient and less imposing valley than Doe Creek, which is a noteworthy feature extending to the divide at the opposite end of the lake. The anticline is more widely breached at the lake than at the divide, where the two ridge crests of Clinch approach each other somewhat more closely.

Sixty-five soundings showing a maximum depth of 75 feet were made, and it is probable that this is approximately the greatest depth, although depths of over 100 feet were reported to the author by local inhabitants, none of whom had ever sounded the lake.

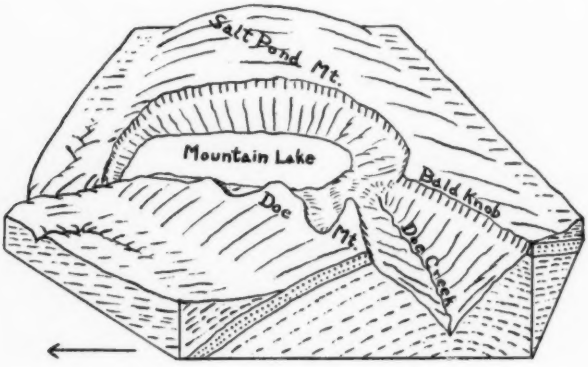

FIG. 2.—Block diagram of Mountain Lake and vicinity

The author's soundings indicate the head of the lake to be shallow, the depth to increase gradually northward reaching 75 feet within a few hundred feet of the outlet, and then to decrease rapidly until the outlet is reached. Near the outlet, within the area of decreasing depth, the bottom is composed of immense blocks of Clinch; the depth here may vary several feet within a horizontal distance of a few inches according to whether the lead strikes or just misses the edge of one of these great blocks. These blocks in the lake are continuous in occurrence with those above water near the eastern side of the outlet, where a veritable rock city of tremendous Clinch blocks is found. Some of these, half submerged, may be dimly seen at the edge of the lake in Figure 1, although their great size is better indicated in Figure 3. These are several hundred feet from the par-

ent ledge forming the wooded mountain ridge. Blocks of importance do not occur west of the outlet, but on the east they can be traced in unbroken procession from ledge to lake bottom.

For the origin of this lake basin several hypotheses may be entertained. If the lake occurred in New York, it would undoubtedly be attributed to glaciation without further study; in that state Lake Minnewaska occupying an anticlinal ripple in the resistant conglomerate of Shawangunk Mountain seems to possess several fea-

FIG. 3.—Clinch blocks situated about midway between parent ledge and lake bottom

tures analogous to Mountain Lake. Minnewaska has been explained as the result of glaciation, but the possibility arises that in part at least it is not glacial, when it is seen that this origin need not be considered for Mountain Lake.

The possibility that the lake occupies a solution basin must be considered; two important limestones, the Shenandoah and the Chickamauga, underlie the Sevier shale in which the lake basin is formed, and the valley areas where these limestones outcrop are pitted with sinks indicating a high degree of solubility. Serious objections to this theory of origin, however, must lead to its rejection. The lake basin is larger than any limestone sink in the area, most of which are small circular basins. Furthermore their bottoms are mostly very uneven, containing minor sinks and knobs, and their

sides are very ragged and upon submergence would form an irregular shoreline. Both of these features are lacking at Mountain Lake where the bottom slope is notably regular, and the shoreline very even. It must also be noted that no sinks in the limestone area contain lakes of any considerable size. Perhaps the strongest evidence that the lake does not occupy a limestone sink is its situation near the top of the Sevier with approximately 1,000 feet of slightly soluble or insoluble shale between it and the top of the limestone. It seems impossible that limestone solution could have any effect at the surface through such a thickness of shale.

All the features noted by the author are consistent with an origin of the lake basin as a normal stream valley subsequently dammed. The longitudinal profile of the basin with its gradual descent down valley to a point of sudden ascent is that of a stream valley dammed by débris, the dam causing the reversal in slope. The genetic history seems to have been as follows: a resequent stream, tributary to Little Stony Creek, flowing down the flank of the Salt Pond Mountain anticline cut through the Clinch forming the mountain arch. With the exposure of the underlying Sevier shales this stream developed a short subsequent anticlinal valley more or less parallel to the trend of the fold and now the site of the lake. At the same time Doe Creek was working headward along the axis of the anticline, and the headwaters of the two streams approaching each other entirely breached the anticline, leaving only a low divide on the Sevier. As longer slopes of Sevier were exposed just south of the notch cut in the Clinch, great blocks of this formation crept downward over the shale slopes, gradually filling the valley bottom just above the notch. The Clinch may also have fallen as talus, or as a rockslide, although there is no evidence of the latter. Weathered particles and organic material filling the interstices made a water-tight dam impounding the lake. The source of the blocks is clearly the ledge above the outlet on the east, and the outlet stream is crowded over to the west edge of the gap with no important blocks farther west of it. The bottom of the gap is paved with smaller irregular Clinch boulders and sand.

Stumps rooted 1 or 2 feet below the surface of the lake indicate a

fairly recent increase in depth. There are no indications that this rise in waterlevel is due to artificial causes, and it may be attributed to a more thorough sealing of the interstices of the block dam.

Comparison of the gradients and the situations of the outlet stream and of Doe Creek indicate that the latter with its much steeper gradient is encroaching upon the area of the former, which has the disadvantage of superposition across the Clinch. Thus the divide, now about 60 feet above the lake and 600 feet southwest of it, is being pushed lakeward by Doe Creek which may eventually tap the lake and establish a new divide at the gap occupied by the present outlet.

THE FORMATION OF GLACIAL TABLES, GRAND TETON NATIONAL PARK, WYOMING

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ABSTRACT

The rock débris which lodges on the surface of mountain glaciers may, on the basis of its history after accumulation, be classified under three heads: (1) small fragments which in summer quickly become heated through and melt themselves down beneath the surface; (2) larger boulders which protect the ice beneath them and so give rise to glacial tables; and (3) blocks of intermediate size which may follow either course (1) or (2) depending upon the extent to which they are heated. The largest number of glacial tables is usually found in early summer, when classes (2) and (3) both give rise to tables.

Unlike glaciers of the ice-cap type, most valley and cliff glaciers carry on their upper surface a scattering of rock débris contributed from the walls of the canyon or cirque in which they are enclosed. Not uncommonly the quantity of such detritus is impressive, even on small cliff glaciers comprising hundreds or thousands of boulders. The history of this rock débris will be the subject of the present and a succeeding paper.

Certain aspects of the subject are well known. By influencing the melting of the ice the rock detritus modifies the topography of the glacier in two ways: (1) small fragments lying on the ice become warmed through and (since the rock absorbs heat more readily than the ice) melt the ice under them and sink beneath the surface; (2) with larger boulders the case is different for though rock is a good absorber of heat it is a poor conductor and consequently the lower surface of such boulders is not notably warmed. Also the ice beneath the boulders is shaded from the sun and is therefore melted more slowly than that around it. Consequently as ablation lowers the surface of the glacier, the boulders become perched on lengthening pillars of ice, and so originate the curious mushroom-like structures to be seen in summer on most mountain glaciers. These "glacial tables" (in some accounts referred to as "ice tables" and "glacier tables") early became familiar through the descriptions and sketches of Tyndall, Forbes, Russell, and other glacialists, and are briefly considered in most textbooks.[1]

[1] E.g., Chamberlin and Salisbury's *Geology*, Vol. I, *Processes* (1909), text on pages 268–70, on which the above paragraph is based.

It is evident that the course followed by each rock fragment—whether it will tend to melt itself down beneath the surface of the glacier or will in time come to stand above it on an ice column—depends principally[2] upon its *size*, and the question arises, What size limit serves as a division between the two possibilities? Or, stating the question differently, What is the *minimum* size for boulders

FIG. 1.—Glacial tables of all sizes in central portion of Teton Glacier, July 7, 1931

capable of forming glacial tables? During July and early August, 1931, the writer sought evidence bearing on these questions in the course of visits to several glaciers in the Teton Range of northwestern Wyoming. Teton Glacier, a typical and accessible ice field on the east slope, offered a particularly excellent field for observation.

On the first 1931 visit to Teton Glacier (July 7) hundreds of glacial tables were found, for the most part clustered in the central and east-central portion of the ice field (Fig. 1). On this date the smaller of the capping boulders were found to be 8 to 10 inches in diameter and

[2] The shape, composition, and texture of boulders are additional factors which to a lesser extent enter into and complicate the question above considered.

of approximately the same thickness.[3] But on the second visit to the same area, eleven days later (July 18), no boulders less than 18 inches in diameter were found standing on ice columns, those of

FIG. 2.—Well-developed glacial table on Teton Glacier, August 3, 1931. The boulder has a maximum diameter of about 10 feet.

smaller dimension which on July 7 formed tables being now sunk into the ice. The change was undoubtedly due to the increased heat of summer, small boulders which before protected the subjacent ice being now heated through and hence caused to melt downward into the ice. Additional proof that small boulders at times form tables and at other times fail to do so was furnished by repeated instances

[3] I. C. Russell, in his account of the glaciers of the Sierra Nevada of California (based on observations made during the summer of 1883) reported that on the Parker Creek Glacier the smallest blocks which gave rise to tables were 16 by 10 by 10 inches (I. C. Russell, *Glaciers of North America* [Boston, 1897], p. 44).

of boulders which had melted down into the tops of broad ice columns to which they themselves had earlier given origin.

On none of the observations made after July 7 were boulders 8 to 18 inches in diameter again found standing on pillars in the central portion of the glacier, but on each visit it was found that near the south margin of the glacier there were many boulders within this

FIG. 3.—Large glacial table in process of formation on Teton Glacier, August 19 1929.

smaller range of size that still formed tables. For example, on July 25 in the central area boulders less than 18 inches in diameter were imbedded in the ice, whereas near the south margin many only 8 or 10 inches in diameter were standing on columns. It was apparent that boulders in the central area, exposed to the direct rays of the sun during the forenoon and early afternoon, became heated through much more effectively than those near the south margin which through the greater portion of each day lay in the shadow of cirque walls.

On the basis of these observations the conclusion is reached that no fixed lower limit of size can be assigned for boulders capable of

forming glacial tables, since the minimum size varies from time to time in a given place and from place to place at a given time. On the one hand is a group including the small rock fragments which tend to melt themselves down beneath the surface of the glacier, and on the other hand is a group including the larger boulders which in time come to stand above the surface as glacial tables; between these groups is a third, comprising boulders of intermediate size, which may follow either course depending upon the extent to which they become heated.

It is evident that when conditions are such as to permit boulders of this intermediate group becoming perched, the total number of glacial tables in a given area will be considerably greater than at times when only the larger boulders become perched. This was strikingly exhibited on Teton Glacier, for it was estimated that on July 7 there were nearly twice as many glacial tables in the central area as on July 18. As a rule, then, the largest number of glacial tables may be found early in the summer, when the heat is not yet so great but that the smaller boulders can form tables. It should be pointed out, however, that the wholesale elimination of these small glacial tables which takes place later in the season is to a greater or less extent compensated for as tables form from the additional large boulders which at this period fall onto the glacier in considerable numbers and from the englacial boulders which are constantly being exposed by melting and accumulate on the surface.[4]

[4] Late in the summer englacial boulders may be seen in all stages of uncovering, and the glacialist must exercise care in distinguishing between englacial boulders being melted out of the ice and superglacial boulders melting into it. With small boulders the distinction is difficult to make, but large boulders found imbedded in the glacier may safely be assigned to the former category. In published accounts, however, large boulders so situated have erroneously been stated to have arrived at their position through downward melting.

SLUMP SCARPS

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ABSTRACT

The subsidence in a lava flow after the feeding source has ceased activity often produces scarps closely resembling true fault scarps. The name "slump scarps" is proposed for such features.

The front edges of lava flows and crater walls are included in Lahee's[2] list of cliffs. Another type of volcanic cliff, produced by subsidence, is frequently found in regions where pahoehoe lava flows are common. Daly[3] mentions constructional lava scarps without definition.

FIG. 1.—Lava bench on ash cliff in Kilauea crater. After rising to the level of this bench, the flow slumped after the feeding source became inactive. Photo by Finch.

[1] Published by permission of the director of the U.S. Geological Survey.

[2] F. H. Lahee, *Field Geology* (3d ed.), p. 279.

[3] R. A. Daly, *Igneous Rocks and Their Origin*, p. 133.

Figure 1 shows a scarp produced by a lava flow from Kilauea in 1921. The branch of the flow that formed the scarp was a small one and the place where the photograph was taken is but 200 or 300 feet from the end of the flow. A more voluminous flow could easily have produced a much larger scarp. After activity at the source of the flow had ceased, the still molten lava under the cooled crust drained out to form tongues at the edges of the flow and the original surface

FIG. 2.—Slump scarp at base of Hat Creek fault scarp. Note spur at left. Photo by Finch.

sank to a much lower level. Cooling and crystallization by decreasing the volume would be but minor contributing factors to the lowering. In this particular case the lava, adhering to an old fault cliff in an ash layer, marked the height of the flow before subsidence. Had there been trees in this locality prior to the flow, their former presence would have been indicated by tree casts.[4]

If the flow is one of great fluidity, the level of the lava surface may lower greatly after the source of supply has given out. The lowering is likely to be appreciable even in very viscous lava, if some obstruction causes the formation of a temporary pool or lake in the flow. If a

[4] R. H. Finch, "Tree Casts and Tree Molds," *Volcano Letter*, January 15, 1931.

flow does not encounter a hill, cliff, or trees, in most cases no evidence will remain to show the amount of lowering of the original surface. A cliff produced by such sinking in a lava flow may appropriately be called a "slump scarp."

In some cases where a youthful flow has produced a slump scarp along the face of an old fault, one easily might be misled into thinking that the slumping was really evidence of recent movement along the old fault. This would be true especially if the flow were masked by sediments. Blackwelder[5] and others have pointed out the difficulty in the recognition of fault scarps. It is easily possible, of course,

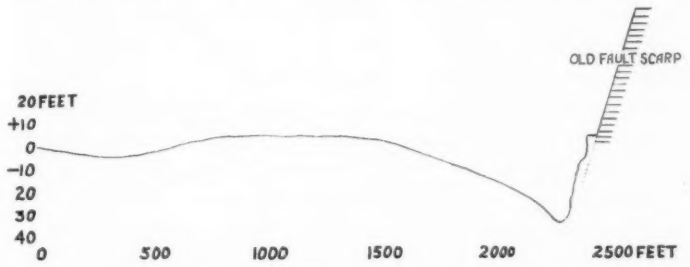

FIG. 3.—Profile across Hat Creek lava flow. Vertical exaggerated ten times. The slump scarp is at the right where it masks the old fault.

to have a combination of slumping and actual deep-seated movement along a fault. It is not uncommon to find places where the evidence is not clear cut. Figure 2 shows a scarp that seems to be due to slumping in a large lava flow from the north slope of the Lassen massif where it butted against the Hat Creek fault scarp. A spur from the old scarp, which is several hundred feet high, made conditions ideal for the development of a lava pool in this vicinity, and hence a scarp of appreciable height. A profile of the flow is shown in Figure 3. It crosses the flow at nearly a right angle and starts from an arbitrary zero at the western edge of the flow.

There are several scarps in a lava flow in Siskiyou County, California, about 17 miles south of Glass Mountain. One, produced where the lava encountered a cone and followed it so as to border an appreciable arc of its base, is clearly due to slumping.

[5] Eliot Blackwelder, "The Recognition of Fault Scarps," *Jour. Geol.*, Vol. XXXVI (1928), pp. 289–312.

FOSSIL FOOTPRINTS IN COMANCHEAN LIMESTONE BEDS, BANDERA COUNTY, TEXAS

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An interesting occurrence of fossil footprints in the bed of Hondo Creek, 26 miles north of the town of Hondo, Texas, was recently brought to my attention.[1] The footprints extend over an area of about 200 feet in length along the creek bottom near the Humble pipe-line crossing on the farm of Al. Mangold, in the Robert M. Williamson Survey, Bandera County, Texas.

The accompanying sketch (Fig. 1) shows the relative position of the footprints and the direction in which the creatures by which they were made, presumably dinosaurs, were moving. There are three distinct sizes of footprints: the largest measures 16 inches across the spread of the toes and 16 inches from the tip of the middle toe to the base of the heel; the next size is 14 by 14 inches; and the small tracks are 10 by 10 inches. The depth of the impressions varies from 1 to 5 inches; the longest pace measures 6 feet, and the shortest pace, 4 feet. In some of the footprints it is possible to observe where the toes gripped the surface over which the animals walked by digging in more deeply and approaching each other at their extremities. An effort was made to detect evidence of the possible drag of the tails of the dinosaurs, but the stream bed is so complexly incised by grooves and other effects of erosion that any possible tail impressions are obliterated.

The limestone in which these footprints occur is a part of the Glen Rose formation of Comanchean age. The bed containing the footprints is itself not fossiliferous, but the immediately overlying bed contains a large assemblage with typical Comanchean forms. Although the footprints themselves are in limestone which must have had the character of ooze when the tracks were made, the overlying

[1] By Mr. Claud Mangum, an employee of the Humble Pipe Line Co.

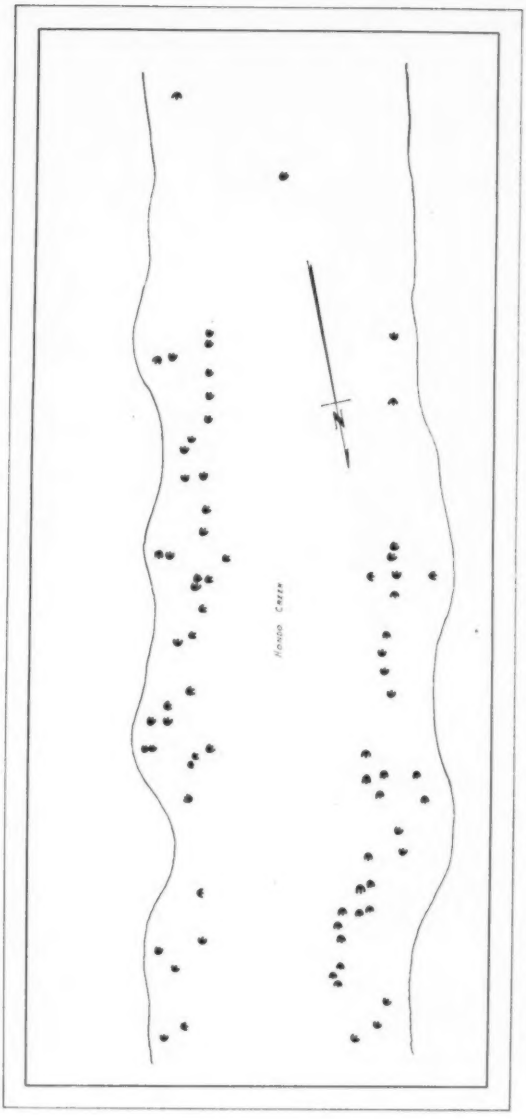

FIG. 1.—Fossil footprints exposed in Hondo Creek, Bandera Co., Tex.

FIG. 2.—Four photographs of the dinosaur footprints from Hondo Creek, Tex.

bed is shaly. Apparently mud was deposited by shallow water in the footprints immediately after they were formed—a circumstance which probably is, in part, responsible for their preservation.

Similar footprints in Hamilton County, Texas, have been described by W. E. Wrather,[2] who in turn refers to a description by E. W. Shuler[3] of dinosaur footprints in Somervell County, Texas. Gould[4] has recently described footprints from several other localities in the Glen Rose formation in Texas.

[2] *Jour. Geol.*, Vol. XXX (1922), pp. 354–60.

[3] *Amer. Jour. Sci.*, Vol. XLIV (1917), pp. 294–98.

[4] *Bull. Geol. Soc. Amer.*, Vol. XL (1929), pp. 458–61.

"WEATHERING OF THE MEDFORD DIABASE—PRE- OR POSTGLACIAL?"

A DISCUSSION

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The article by Mr. Arthur Wolf, originally published in the *Journal of Geology*,[1] and presented by the author and Professor A. C. Lane at the meetings of the Geological Society of America in December, 1932, favors the theory that the conspicuous weathering shown by the Medford diabase of eastern Massachusetts took place before the advance of the last ice sheet over the Boston region. We do not believe that the author has adequately presented the data and we cannot agree that his interpretation of the evidence is correct. We present in this note what we believe to be a more consistent interpretation of his evidence and in addition we shall introduce other data which, in our opinions, refute his conclusions. We believe that the observed weathering has taken place since the last ice invasion.

The weathering of the Medford diabase may be characterized as spheroidal or concentric. Merrill[2] describes it as follows: "The rock has undergone extensive disintegration, giving rise to loose sand and gravel of a deep brown color, in which lie rounded boulders of all sizes of the still undecomposed material. These boulders, as is usually the case, show a more or less concentric structure, from without inward, until a core of unaltered diabase is met with." The weathering started along the joints and has proceeded inward to varying distances to leave only rounded cores of the once angular blocks.

Neither the author nor any previous writer has offered any explanation of the weathering. The process is mainly mechanical disintegration, and a possible explanation is that a slight amount of

[1] Vol. XL (1932), pp. 459–65.

[2] George P. Merrill, *Rocks, Rock-Weathering, and Soils* (New York: Macmillan Co., 1897), p. 218.

hydration of the ferromagnesian minerals causes an increase in volume which results in a mechanical breakdown of the rock.[3] Once the diabase has disintegrated into small units, chemical weathering may take place fairly rapidly.

In the following discussion we shall first consider the arguments used by the author, as far as possible following the order which he has employed; then we shall present new evidence which we have obtained.

DISCUSSION OF EVIDENCE CITED BY AUTHOR

1. *Chemistry of the weathering.*—The author presents a table[4] which shows analyses of both fresh and altered material by Whitney and Merrill, and one analysis of fresh material by Hobbs. Unfortunately all but one of the six analyses have errors, either in the addition or in quoting data published elsewhere. The errors are not serious, but we should like to present a more detailed interpretation of these analyses. The author states:[5] "If the disintegration was caused by the leaching effect of the postglacial waters from the melting ice or from a wet postglacial climate, it seems highly probable that the weathered diabase would show a large decrease in its more soluble constituents such as the soda, potash, and silica, and a correspondingly large increase in the alumina, the most insoluble constituent. But as the analyses show, there is comparatively little attack on the feldspathic constituents; the alumina shows a slight increase to be sure, but the soda also unaccountably increases, while the potash decreases very little." Merrill[6] in his consideration of such analyses states: "We must first of all remember that any indicated loss or gain of a constituent may be only apparent, and that the true relative proportions can be learned only by calculating results of analyses of both fresh and decomposed materials on a common basis." Continuing he says: "Of all the essential constituents occurring in appreciable quantities in siliceous crystalline rocks the alumina and iron oxides are the most refractory and the least liable to be removed by a leaching process, although they may undergo manifold changes in

[3] Eliot Blackwelder, "Exfoliation as a Phase of Rock Weathering," *Jour. Geol.*, Vol. XXXIII (1925), pp. 793–806.

[4] *Op. cit.*, p. 460.

[5] *Ibid.*

[6] *Op. cit.*, p. 208.

mode of combination." With this last statement the author is in agreement when he points out that he would expect an increase in alumina in the analysis of leached diabase.

Although we recognize that there may be some slight error in assuming Al_2O_3 to remain constant we feel that it is the nearest possible approach to the truth. Following the methods described by Merrill[7] and using the analyses given in Mr. Wolf's paper, we have calculated the percentage of each constituent lost in the weathering (Table I).

TABLE I

PERCENTAGE OF EACH CONSTITUENT LOST IN WEATHERING OF MEDFORD DIABASE

	A	B	C
SiO_2	18.03	11.74	2.88
Al_2O_3	0.00	0.00	0.00
Fe_2O_3	18.10	gain	gain
FeO		46.42	50.48
CaO	25.89	19.61	39.87
MgO	21.70	29.89	37.26
MnO	41.57	33.19	41.41
K_2O	29.15	40.01	24.10
Na_2O	12.83	6.90	gain
P_2O_5	11.39	gain	gain
H_2O	gain	gain	gain

A. Based on analysis of fresh and weathered rock by G. P. Merrill (*Rocks, Rock-weathering, and Soils*, p. 221).

B. Based on analysis of fresh rock No. 6 and weathered rock No. 5 by C. M. Whitney.

C. Based on analysis of fresh rock No. 6 and weathered rock No. 2 by C. M. Whitney. Analysis of No. 2 of Whitney's totals only 98.15 per cent and cannot be considered a first-class analysis.

This table demonstrates that a large percentage of the soluble constituents have been lost and the author's statement that "CaO, MgO, and MnO decrease a trifle, which is in accord with a slight leaching action," is incorrect. The contention that the chief chemical change has been oxidation with little leaching is thus unsupported.

By concluding that the type of weathering shown by the Medford diabase could not be produced in the "wet postglacial climate" the writer implies, although he does not say so explicitly, that the weathering took place in a drier climate, possibly a semi-arid climate.

[7] *Ibid.*, p. 210 n.

There is no collateral evidence for such conditions, and thus no reason for believing that the Medford type of weathering could be produced more readily in preglacial than in postglacial time.

2. *Depth of weathering.*—We question the accuracy of the author's statement that complete disintegration has locally progressed to a depth of 40 feet. It is true that along Governor's Avenue, Medford, there is a 30° slope, approximately 40 feet high, on which disintegrated diabase may be found from top to bottom (Fig. 1 of the author's paper). It is evident in the field as well as in the author's picture that there are ledges of unweathered diabase at or very near the surface of this hillside. The weathering reached a maximum exposed depth of 5 or 10 feet along joints, but in places the unweathered diabase forms the bed-rock surface directly under the glacial till. We believe that the weathering here is a surface phenomenon extending downward only along fractured or porous zones. The exposure of the material on this steep slope evidently led the author to conclude that the depth of weathering is equal to the height of the hill. Can the author cite an example of 40 feet of complete disintegration in a vertical exposure?

Even if the depth of disintegration were 40 feet, we doubt that it would necessarily prove or suggest that the weathering is older than the last ice invasion. In fact, under favorable conditions weathering may have progressed to scores of feet in postglacial time. This is particularly true of coarse-grained rocks cut by numerous joints and exposed on steep slopes. Water may move down joint planes to depths of many scores of feet beneath the surface and then emerge at the base of the slope. Hydration and other weathering processes may take place all along the joint and thus the rock be rapidly weathered to great depths. If joints and other fractures are closely spaced, a large mass of rock may rapidly disintegrate. In Franconia Notch, New Hampshire, to be described in greater detail on a subsequent page, we believe that post-Wisconsin disintegration has progressed to depths of at least 30 feet. We wish to make it clear, therefore, that, although we doubt that the weathering of the Medford diabase extends to such great depths as the author states, even if it did it would not be positive evidence for his thesis.

3. *Weathered boulders in till.*—Figure 3 of the author's paper shows

a diabase boulder about 16 inches in diameter imbedded in glacial till which contains a large mass of varved clay. Although this exposure is no longer available we visited the locality when excavations were in progress and are thus familiar with the field relations. The boulder now consists of a solid, fresh core surrounded by a shell of completely disintegrated diabase. We have difficulty in reconciling these facts with the author's thesis of preglacial weathering. There seems to be only one logical explanation of the observed facts, namely, that a relatively fresh boulder of diabase was picked up, transported at least a short distance by the ice, and then dropped into the till. Since that time an outer shell several inches thick has completely disintegrated. The author may suggest that a diabase boulder underwent partial disintegration before the last ice invasion, was frozen, picked up by the ice, and finally incorporated in the till. The boulder would certainly have to be treated with the utmost discretion by the ice to prevent removal of the shell on one or more sides. Other boulders of diabase show similar shells but of varying thickness. This is quite compatible with a hypothesis of postglacial disintegration, for weathering is notoriously erratic in its attack. It would be entirely possible to find disintegrated and relatively fresh boulders very close to one another in the same exposure of till, even if the weathering were postglacial.

Figure 2 of the author's paper leaves much to be desired in the matter of clarity. In the foreground there is a mass of quarried blocks and in the background what appears to be a large boulder overlain by glacial till or weathered diabase. The fresh and weathered boulders mentioned are not designated, nor is the till, and since the exposure no longer exists, we were unable to ascertain the facts. We wish to question chiefly the author's explanation of the relationships shown. He contends that the only explanation is that two of the boulders were weathered before they were picked up by the ice and the third was a fresh boulder. The author has failed again to recognize the difficulty of transporting rotten boulders by glacial ice without partial or complete destruction. Further, he has failed to recognize one of the outstanding characteristics of weathering, namely, that it can be and often is highly differential. The present writers believe that it is possible to have such differential weathering in glacial till. Completely weathered boulders in the till may have been

slightly attacked before the last ice invasion, but it is highly improbable that they were completely disintegrated. The dominant part of the observed weathering of necessity must have been postglacial.

4. *Freshness of till.*—In two places the author refers to the glacial till associated with the weathered diabase as "relatively fresh." In the abstract for the Cambridge meeting of the Geological Society of America[8] the fact that the till is "fresh" is emphasized: "Fresh till enclosing diabase pebbles in various stages of decomposition; fresh till overlying deeply disintegrated reddened diabase." It is not clear, however, exactly what is meant by "fresh till" and what their concept of weathered till would be. In all cases observed by us the till is invariably yellow, indicating some oxidation. Apparently, however, because it is not stained so deep a brown as the disintegrated diabase the author concludes that the till has not been exposed to weathering agents as long as the diabase. This is not a legitimate comparison, however, for in a given interval of weathering a diabase with 15 per cent iron oxide will obviously be stained a much deeper brown than a till with a low content of iron. Excavations in and near Boston have shown that the fresh till is bluish-gray, whereas the till nearer the surface is yellow, presumably due to oxidation. Hence, to call yellow till "fresh" does not seem justified.

5. *Rate of weathering.*—The author concludes his argument with a discussion of the rate of modern disintegration of the diabase. We quote: "For evidence of the rate of weathering a historical cemetery on Salem Street in Medford Square, Medford, was carefully investigated." We can conceive of no excuse for comparing the rate of weathering of a rock in natural surroundings with a slab of the same rock chosen for monumental purposes because of its lack of joints or cracks, smoothed or polished, and then stood on end, thereby removing it from possible attack by ground water charged with organic acids.

ADDITIONAL EVIDENCE

Additional field work has brought to light some new and interesting evidence on this problem.

[8] Alfred C. Lane and A. Wolf, "Weathering of the Medford Diabase," *Bull. Geol. Soc. Amer.*, Vol. XLIV (1933) (abstract), p. 92.

1. In front of Forest Park School, opposite 204 Governor's Avenue, Medford, there is an excellent example of glacial striations on diabase *in situ*. Construction work has exposed the rock to a depth of $2\frac{1}{2}$ or 3 feet. This outcrop on top of a hill was fully exposed to the advancing ice, yet it is sufficiently disintegrated so that one may readily pick it apart with a prospector's pick. On the striated surface, contraction cracks due to weathering have developed, which are certainly poststriation in age. There is only one logical interpretation of this evidence. The diabase was a fresh, hard rock when the ice passed over it and the observed disintegration has taken place since that time. Another striated surface was observed at the end of Pilgrim Road, Medford; the disintegration is more pronounced and extends to an observed depth of 3 feet with no deeper exposures.

2. In the exposure shown by the author (Fig. 1) is a bit of evidence overlooked by him, which we believe has a damaging effect on his thesis. In the diabase here are numerous fractures along which there has been considerable silicification and in places druses of quartz crystals are observed. These silicified zones are more resistant to weathering than the normal diabase on either side. At one place where the diabase is completely disintegrated to a depth of about $2\frac{1}{2}$ feet, two of these silicified zones, $1\frac{1}{2}$ and $\frac{3}{4}$ inches wide, respectively, extend upward through the weathered material to the contact of the diabase and the glacial till which here has an exposed thickness of 3 feet. Evidently the ice at one time rested directly on the surface of the diabase at this point, and if the disintegration were preglacial it seems inevitable that the weight of the ice would have caused compaction of the weathered material if indeed it did not effect complete removal of that material. During the process of compaction these thin silicified zones would have been broken and disarranged. Such is not the case; therefore it seems much more likely that the weathering is postglacial. We believe that the disintegration shown in this exposure is due to water moving down the steep slope of the hill along the contact of the till and the diabase, attacking the diabase along joints and shear zones.

3. We believe that a rapid rate of postglacial weathering is well demonstrated by the biotite granites of the White Mountain region.

Here, at elevations as low as 600 or 700 feet above sea level these rocks show extensive disintegration. Glacial erratics, which were originally as much as 3 feet in diameter, have completely broken down into heaps of arkosic sand. Glacially plucked slopes are covered by disintegrated granite. In Franconia Notch, a glacial trough which has been overdeepened by ice scour, the writers have seen a coarse, biotite granite disintegrated to a depth of 30 feet. This material is well exposed in an excavation for road material. In our opinion the evidence indicates that, under favorable conditions, disintegration of granitoid rocks has progressed to a depth of many feet since the last glacial advance.

In conclusion, we agree with Crosby, Barton, and Merrill that the evidence is conclusive that the observed weathering of the Medford diabase is younger than the last ice invasion.

REPLY

ALFRED C. LANE AND ARTHUR WOLF
Tufts College

We are much obliged to the *Journal of Geology* and the authors Billings and Roy for the chance to record our side of the discussion of the Medford diabase. We will follow the order of presentation in their article.

As to the chemistry, unfortunately, G. P. Merrill and C. N. Whitney are both dead and we cannot profitably discuss errors in the analyses. Billings and Roy say they are not serious and use the analyses in preparing Table I. But we do find an error in the reference to the thesis by C. N. Whitney, which was a Tufts Engineering Thesis, No. 252, in 1911 and not 1916. None of the analyses would be passed by Washington as A.

We have checked the analyses as published in the *Journal of Geology* with the chart of the analyses in this thesis. They were at the time submitted to G. P. Merrill and we find no errors in the copying of the analyses from the thesis except in the footing of the analysis by Sweetzer from Hobbes where the third figure should be 0 instead of 1. However, our contention is that the weathering is marked by oxidation and hydration more than by the kaolinization of the

feldspar which is still recognizable and has well defined cleavage even in the disintegrated diabase. We think it is clear both in the analyses and in thin section, as well as by ocular inspection, that such is the type of weathering.

We are not quite sure that we understand the authors' objection, because they say[1] that "the process" of weathering "is mainly mechanical disintegration" and that "a possible explanation is that a slight amount of hydration of ferruginous mineral causes an increase in volume which results in a mechanical breakdown. Once this diabase has disintegrated into small units chemical weathering may take place."

To this we agree, except that the ferromagnesian minerals are not only hydrated but oxidized, very probably almost at the same time. That the labradorite, which is one of the chief constituents, is relatively untouched appears from their Table I, where the soda (Na_2O) shows relatively little loss or gain. The same table shows that the iron which changes to iron oxide seems also to increase. This table well shows that the percentages of lime, manganese, and magnesia which we may attribute to the pyroxenes show the greater losses. Here we agree. It would be desirable to get the weight per unit volume of the much disintegrated diabase and check up on the probable real loss. This has been tried by the paraffin dip method. The results have not yet been satisfactory.

At the discussion in Cambridge at Christmas it was suggested by Professor F. K. Morris that the disintegration was a spontaneous affair. We would not deny the possibility of molecular instability and of spontaneous molecular rearrangement in rocks, but we hardly see how the oxidation shown (to say nothing of the losses which Billings and Roy estimate) can exist without the co-operation of circulation from the atmosphere.

We do not say and we do not mean to imply that interglacial climates were necessarily different from those more recent, but it does seem to us likely that the conditions of the deep disintegration were such[2] that oxidation could penetrate more deeply than the present

[1] Pp. 654–5.

[2] Compare, for instance, the footnote to page 48 of G. Russell Smith's *Industrial and Commercial Geography*, on Marbut's work, and P. E. Raymond, *Amer. Jour. Sci.*, Vol. XIII (1927), p. 234.

effect on till and present weather indicate, especially as the diabase is in places covered with rather impervious till. And the present, postglacial weathering seems to whiten the diabase (kaolinize it?) and attack the feldspar more than this earlier weathering.

The exposure mentioned as "new and interesting evidence" near the Forest Park School has been visited twice a year, since 1927, by the Tufts classes. The striated surface is whitened, since striation and this postglacial weathering has seemed to us to go down only about ½ inch,[3] whereas the brown disintegration goes much deeper.

The owner of part of the land on which the Forest Park School stands dug deep pits and sold the decomposed gravel therefrom, and, we are told, claimed that it was 40 feet deep. Had it been realized that the depth of weathering recognized and apparently agreed to, both by G. P. Merrill and A. W. G. Wilson,[4] as likely to be between 30 and 50 feet, would be questioned, it would have been well to refer to a threatened lawsuit over the land of the Lake Forest School. The natural surface there is reasonably flat and not subject to the side hill uncertainty of the exposures south.[5]

Accordingly, a series of borings were made in 1926 by B. F. Smith and Company, and a report made by C. A. Farwell, and the records are in the office of Fay, Spofford, and Thorndike; and Lane was consulted at the time. The records show that Wilson's Figures 2 and 3[6] are excellent diagrams of the conditions. The undisintegrated diabase was reached at quite variable depths, but in some cases not even at 20 feet.

It is quite clear from Wilson's Figures 2 and 3 that when he alluded to "fifty feet"[7] he was not referring to any complete disintegration! Nor was Merrill when he referred to "30 or perhaps 50 feet."[8]

It may be said, however, that Section III of Wilson's Figure 3 shows the Governor's Avenue cut, and that, when the street was

[3] A. W. G. Wilson, "The Medford Dike Area," *Boston Soc. Nat. Hist.*, Vol. XXX, No. 2 (1901), near bottom of p. 366.

[4] *Ibid.*, pp. 366–68.

[5] *Ibid.*, Sec. III, Fig. 1, p. 361. The landowner claimed that the "red gravel" produced by the disintegrated diabase was worth $2.00 a load and that there were 12,000 tons of it!

[6] *Ibid.*, pp. 367–68.

[7] *Ibid.*, p. 368.

[8] *Bull. Geol. Soc. Amer.* (1896), p. 359.

first cut through and visited by Lane and others, there were naturally better exposures than at present. At that time it was still not quite down to the bottom of the zone of disintegration.

What Wilson says[9] as to the planing off and striation of a surface containing residual boulders is excellent, and it was not thought necessary to repeat it.

One other Tufts Chemical Thesis, No. 29, "The Chemical Examination of Soil with Reference to Geological Formation and Crop Raising," by Arthur Welch Philip and Ralph Oliver Phillips, contains a partial analysis of the Medford diabase soil, but adds little positive information as the oxides of iron were not separated.

About a hundred yards from the Forest Park School there is a cellar hole over 8 feet deep which shows a section of the glacial deposits underlain, down as far as it goes, by decomposed diabase. This place, which we are sorry we did not mention in the paper, is one to which we invited a trip at the time of the Boston meeting of the Geological Society of America. It is one of a number of places where glacial deposits and the underlying weathering are clearly shown. There is probably still visible another place on Circle Street nearby.

We do not, however, imagine and did not mean to imply that anywhere we have seen "40 feet of complete disintegration." Nowhere has it progressed uniformly to anything like 40 feet. In our figure at Governor's Avenue, boulders of relatively fresh diabase can be seen high up on the slope, and the lower part of the slope is but a talus of the fine disintegration produced. We do not think that any part that we have seen was necessarily 40 feet from the original bed surface although Governor's Avenue has cut in somewhat to the side of the hill.

It is also true that relatively unweathered diabase may be found beneath the till, but there have been exposures with much more than 5–10 feet of disintegration along joints.

We also agree that even a depth of disintegration of 40 feet does not necessarily prove that the weathering is older than the last ice invasion, but we do think it suggests it. The depth of disintegration is more like that around Washington, D.C., than that normal to areas recently left by the Wisconsin glacier.

[9] *Op. cit.*, pp. 366–70.

We do not take up the deep weathering which Lane and Keith have seen near Mt. Vernon, New Hampshire, and suspect to be pre-Wisconsin, or other occurrences such as the Blandford kaolin described by W. O. Crosby,[10] because each case of weathering is really independent, and should be so handled. But as Billings and Roy have mentioned New Hampshire occurrences, we may suggest that a glacier is like a river of ice and like a river of water cuts in some places and deposits in others, that immediately to the north is the higher land of the Middlesex Fells, and to the south the group of deposited drumlins of Somerville, so that the Medford diabase is located in a place where it would be quite natural that the last ice advance should not remove all previous weathering.

With regard to Figure 3, exposed during the excavation of the Cousins Gymnasium, it is a pity that the photograph does not bring out well enough what seems to us clear, that up in the till was a rounded residual boulder, nugget, or cannon ball such as were numerous in this excavation and in general visible when one works from the rotten diabase surface down into the more solid parts. We have other good photographs which we should be glad to make available, and, thanks to Professor Munro, excellent photographs of exposures no longer visible. Unfortunately they are not color photographs.

We see no more difficulty in the ice transporting boulders in various stages of decomposition than in transporting the mass of varved clay shown in Figure 3, which we all saw. That varved clay was not yellow to our eye, nor was the till of other exposures we have seen of other colors than the till of fresh exposures around Boston, except that when it overlies the rotten diabase it incorporates more of it at the bottom. In fact, Powers suggested that some of the lower part might be an older (Illinoian?) till. We might have saved samples of that mass of varved clay and had analyses to compare with analyses of the brick clay made by L. H. Carter in his thesis and quoted by Lane in his paper.[11]

These brick clays have very much more ferrous iron in proportion to ferric iron than the weathered diabase, and it should not be diffi-

[10] *Technology Quarterly*, Vol. III, p. 28.

[11] *Geol. Soc. Amer. Bull.*, Vol. XXXIII (1922), pp. 363–69.

cult to prove the fact chemically. Of course, any exposure of till rapidly changes somewhat in color.

It would be an excellent subject for theses investigation to get the relative proportions of ferrous and ferric iron in various apparently more or less fresh glacial deposits, especially where they directly overlie the decayed diabase.

The little silicified vein in Figure 1 was by no means overlooked. Exactly comparable quartz veins are found in the disintegrated rock around Washington. It is one of the clearest proofs that the rock has rotted in place, has not been disturbed, and, as it stops at the old surface, shows very well just where the till begins.

But that it should be disturbed by compaction is not at all likely when we agree that there is a nexus of cannon balls of solid diabase to support the weight, while in the weathering they might tend if anything to expand.

Finally, as to rate of weathering,[12] the excuse for choosing only one case (that in the cemetery) was that we did not wish to burden the *Journal of Geology* with multiple instances, but took one likely to be accessible, comparable, and undisturbed. The water tables of Ballou and of the Goddard Gymnasium might have been shown, or the front steps of the Barnum Museum which are seldom used. In all these the weathering on the surface of the diabase of the last fifty to one hundred years is visible, but shallow and of a different color—gray rather than yellow-brown. This gray weathering is practically universal under stripped soil or vegetation. We might also mention a large erratic perched on the top of Pine Hill, and at fit seasons of the year some of the Department are sure to be able and glad to be able to show whatever exposures may be available.

[12] *Ibid.*, Vol. XXIII (1912), p. 292.

REVIEWS

The Story of a Billion Years. By WILLIAM O. HOTCHKISS. Baltimore: The Williams and Wilkins Company in co-operation with the Century of Progress Exposition. Pp. 137; figs. 8. $1.00.

As a part of the program of the Century of Progress Exposition a series of about twenty books has been prepared to present graphically the tremendous scientific advance made in the last hundred years—an advance unparalleled in human annals. Within this span of time has grown up nearly all of our science of geology and to portray this deciphering of the events of more than a billion years in a manner easily understood by the intelligent layman has been the aim of Dr. Hotchkiss. This has been successfully accomplished and the teacher of geology will find stimulation in the technique of presentation. Particularly valuable, it seems to the reviewer, is the realization which the lay reader will get of the great importance to our civilization of geologic resources and geologic investigations which previously, perhaps, he has little suspected.

R. T. C.

The Geological Society of America 1888–1930. A Chapter in Earth Science History. By HERMAN LE ROY FAIRCHILD. New York: Published by the Society, 1932. Pp. 232; portraits 17. $2.50.

The formative stages of the Geological Society of America have now been chronicled by one of its founders. Fortunate the Society is in having this done by Professor Fairchild, who is not only the sole survivor of that small group present at the first annual meeting at Cornell University in 1888, but who later served the Society as secretary for sixteen years (1891–1906) and as president in 1912. No one is more familiar with the Society's inception, early struggles, and enlarging activities than he, and probably no one else is so well qualified to throw its development into true perspective.

In this history we read of the dawn of earth science in the Greek and Roman world, of its long eclipse by the heavy clouds of medieval darkness, and its gradual emergence with the increasing light of modern times. Chapters on "Geology in Western Europe," on "Geology in America before 1848," and "Geology in America, 1848–1888" prepare the way for the

genealogy of the present society, whose clear and honorable lineal descent Professor Fairchild traces through four generations. The great-great grandparent was the Association of American Geologists, organized in 1840; the great grandparent was the Association of American Geologists and Naturalists, dated 1843; the grandparent is the yet flourishing American Association for the Advancement of Science, organized in 1848; and the immediate progenitor is Section E of the latter which dates from 1882. "The Geological Society, the fifth generation in a time space of forty-eight years, is a reversion to type."

The organization of the Society, its early membership, and early meetings are naturally foremost in historical interest, but they have not been given undue prominence. The reader will find a well-balanced, authoritative account of the development of the Society in its different aspects down to 1930. Many well-known names appear and reappear in these pages. Though the succession of presidents has been long and eminent, in its whole history the Society has had but four secretaries (Stevenson, Fairchild, Hovey, and Berkey), but four treasurers (Williams, White, Clark, and Mathews), and but two editors (McGee and Stanley-Brown). These men have carried its heaviest burdens.

The two concluding chapters of this admirable history are: "Geology Since 1888 and Present Status," and "The Look Ahead."

R. T. C.

Report on the Geology of the Ruhuhu Coalfields. By G. M. STOCKLEY and F. OATES. Geological Survey Department, Tanganyika Territory, Bulletin No. 2. Pp. 68; pls. 7; sections; maps 2.

The Ruhuhu coalfields lie in Southern Tanganyika just east of Lake Nyassa. This coal, like most of that in Africa, occurs in beds of Permian age, belonging to the Karoo system. The Karoo is here subdivided into eight divisions with a total thickness of about 5,900 feet.

The Karoo rocks are not widespread in Tanganyika and have been preserved in this area by great downfaulting movements of the Rift Valley system. Thus the fields are bounded by straight-line faults with downthrows of from 4,000 to 10,000 feet which have let the Karoo rocks down against beds of the Primitive System. The tectonics of the area support the tensional theory of the formation of the rift valleys.

There are seven coal fields, four of which have potentialities. The coal seams have thicknesses up to 18 feet. The coal is bituminous, non-coking, and has a fairly high ash content. In view of the probable economic

development of Central Africa and the fact that all coal must be imported from Natal at present, these fields constitute a valuable resource.

Useful addenda to the report are a colored geological map, a sheet of stratigraphic columns and cross sections, and a correlation table of the Karoo rocks of this area with other parts of South Africa and with rocks of similar age in India and Europe.

CHRISTOPHER RILEY

Die schweizerische Eisenerzeugung, ihre Geschichte und wirtschaftliche Bedeutung. By HANS FEHLMANN. "Die Eisen- und Manganerze der Schweiz: Beiträge zur Geologie der Schweiz, Geotechnische Serie." Lieferung XIII, Band 3. Bern: Hans Huber, 1932. Pp. 255; figs. 76.

This book is one of a series dealing with the development of the mineral resources of Switzerland. In the present volume the production of iron is approached from the historical point of view. The account begins in ancient times with the Iron Age in the Mediterranean region, but the setting soon shifts to Switzerland, and other lands are never entirely forgotten.

The first three divisions of the book correspond to three stages in the development of the treatment of iron ore. The substitution of the blast furnace for the bloomery hearth ended the first period. The second period, when only charcoal was available for the blast furnaces, gave way to the third when the use of coke blast furnaces and electric furnaces was made possible by the building of the railroads. Included in these sections are discussions of the mining districts, methods of mining and metallurgy, products, social and legal status of the industry and the workers, and economic matters, including production, consumption, and commerce. The fourth section of the book describes the present condition of the Swiss industry and includes summaries of the reserves and analyses of the ores.

EDWARD L. TULLIS

German-English Geological Terminology—Englisch-Deutsche Geologisch-Mineralogisch Terminologie. By. W. R. JONES and A. CISSARY. London: Thomas Murby & Co.; New York: D. Van Nostrand Co., 1932. Pp. xvii+250. 12*s.* 6*d.*

This book is essentially a textbook in German-English geological reading and as such fits its purpose splendidly. The tables and indexes are so

arranged that the volume may be used as a dictionary as well. Appendixes give the chemical elements, mineral and rock names in both languages, the abbreviations frequently used, and tables of weights and measures. The book may be recommended to those interested in familiarizing themselves with the general terminology of geology in these two languages.

H. W. STRALEY, III

The Miami-Picher Zinc-Lead District, Oklahoma. By SAMUEL WEIDMAN. Oklahoma Geological Survey, Bulletin No. 56 (1932). Pp. 168; tables 26; pls. 11; figs. 12; Bibliography. With chapters on "Mining Methods," by C. F. Williams, and "Milling in the Tri-State District," by Carl O. Anderson.

For half a century the lead and zinc deposits of the Mississippi Valley region have been the objects of periodic study by the federal survey, the surveys of the states concerned, and by others, but many features of their origin still remain obscure. Over most of this period geologic opinion has favored the view that they were deposited by solutions of meteoric origin, but of late years there has been a decided swing toward the view that the ultimate sources of the metals were deep-lying magmas. Professor Weidman, whose report forms the subject of this review, is the latest protagonist of magmatic derivation.

In his folio on the Joplin district, Siebenthal emphasized the concept that the breccias of the "broken ground" mineralization were the result of collapse due to underground solution incident to the development of sink-hole topography on the surface of the Boone limestone. Weidman sees no evidence that a sink-hole topography developed on the Boone surface prior to the deposition of either the Mayes or the Cherokee formations and believes that "variations in thickness of the shale and sandstone of the Mayes formation appear to be the basis to a large extent for the erroneous belief in the widespread occurrence of 'shale holes' and shale depressions in the surface of the Boone formation."

The cherts of the region are interpreted as of two ages. The white cherts, so abundant in the undisturbed portions of the Boone and as fragments in the ore breccias, are believed to have formed in the interval between the Boone and the overlying Mayes deposition, this conclusion being based on the presence of pebbles of these cherts in the basal conglomeratic beds of the Mayes and their absence at higher horizons. They are interpreted as replacements through the agency of thermal solutions coming from buried intrusives, but not accompanied by sulphide mineraliza-

tion. The darker cherts, termed jasperoid, forming a component of the matrix of the ore-bearing breccias, are clearly younger and are interpreted also as of hydrothermal origin, though in this case chertification was accompanied by sulphide mineralization. Thus two periods of hydrothermal action are postulated, one in Mississippian times and one in the Mesozoic. To the reviewer much more evidence seems necessary to establish a hydrothermal origin for the earlier cherts.

Dolomitization is uncommon in the non-mineralized portions of the Boone and Mayes formations, but is common in the mineralized area both in open spaces and in the matrix of breccias; it is believed to have accompanied the sulphide mineralization.

In considering the genesis of the ores the occurrence of post-Carboniferous intrusives at the Rose Dome, Kansas, and Spavinaw, Oklahoma, are cited. Of the intrusive origin of the acid and basic rocks at Rose Dome there can no longer be any question, but at Spavinaw an intrusive origin has not as yet been certainly established. The cryptovolcanic area at Decaturville is cited as belonging to the same category, but Dake and Bridge have recently shown that the intrusion at this place is of Cambrian age. In 1930 granite was struck in a well in the Bird Dog mine near Picher about 500–700 feet above the general level of the pre-Cambrian granite floor of the district. This granite Weidman interprets as intrusive on the basis of its form—of silicification and some arching of overlying sandstones—and the presence in the granite of chlorite, hydrargillite, and epidote which he regards as endomorphic effects of the intrusion. While it is entirely possible that the granite is intrusive, it should be pointed out that although all these minerals may develop hydrothermally they also are common products of weathering.

The granite at the Bird Dog mine is regarded as the immediate source of the ores in that vicinity, some of them only 1,000 feet away. To the reviewer it seems unnecessary from the standpoint of a magmatic hypothesis to postulate so proximate a source for the mineralizing solutions in the total absence of any high-temperature minerals in the ores. It may be that the ore-depositing solutions and granite dike both came from a much more remote magmatic source.

The data which Dr. Weidman has presented form a most valuable contribution and are supplemented by excellent chapters on "Mining Methods," by C. F. Williams of the U.S. Geological Survey, and on "Milling in the Tri-State District," by Carl O. Anderson of the Century Zinc Company.

E. S. B.

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